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THE STRUCTURES OF IRON-ORE DEPOSITS[1]

by

Ya. N. Belevtsev

on-ore deposits — metamorphogenic, n and hydrothermal — have varied geologic ctures, and the emplacement of their ore es usually depends on the combined action e-ore, contemporaneous and post-ore nic movements. For this reason the structure of a deposit must be understood as a com- ion of such elements as determined the lacement, conditions of occurrence and phology of the ore bodies during the course eir development.

thorough study of the geologic structure a correct understanding of its history will le the geologist to estimate the ore pros- s of a deposit and to survey it with the least ible expenditure of time and effort.

the Soviet Union, China, Canada, Sweden most other countries with a developed iron- industry, the overwhelming majority of ores are extracted from Precambrian de- ts of metamorphogenic and complex genesis, ell as from contact-metasomatic and sedi- tary, primarily marine, deposits. Deposits her genetic types, both in the U. S. S. R. and e world in general, are of subordinate im- ance. Hence this article will consider ly the structures of the deposits and ore s of these high-producing genetic types, will touch only in part on the others, which of less industrial importance.

n the basis of the definition given above of concept of the structure of a deposit, we l consider only those elements which have ome degree affected the localization, the lacement and the morphology of the ore es.

The structures of endogenic iron-ore de- its are usually more-or-less complex, reas the structures of sedimentary marine osits are very simple and require no special ussion.

Metamorphogenic and contact-metasomatic deposits are usually emplaced in complex fold-fracture or fold-fault structures in which the ore columns, dislocated stratified occurrences, stocks, lenses and ore nodes have been formed. Deposits of early magmatic and hydrothermal genesis are commonly associated with fault or fault-fracture structures, which contain the dike-shaped bodies, veins, ore nodes and stocks of disseminated ores.

In deposits of these types one almost always can with fair certainty distinguish the elements of the pre-ore contemporaneous and post-ore structure. The structures of sedimentary deposits are usually simple and formed at a single time, although post-ore disjunctive dislocations sometimes occur among them.

A classification of the structures of the principal industrial iron-ore deposits will be found in Table 1 of this article.

I. DEPOSITS OF METAMORPHOGENIC AND COMPLEX GENESIS (KRIVOY ROG TYPE)

These deposits are characterized by three principal types of structures.

1. Fold-fracture structures: in these the ore occurrences of the deposit are associated with large synclinal or, more rarely, anticlinal folds, and the actual producing levels of ferruginous silicate rocks are located in a dense net of various fractures. The latter occur both as small fractures and microscopic cleavage fractures developed within individual thin layers of the rock, and as large bedding-plane fractures and joints which may often be traced for distances of tens of meters or more. The joints are usually open, but there is sometimes a small displacement of the ore along them, attributed to post-ore formation. The ore here almost always contains small-scale folding or plications.

The varieties or subtypes of fold-fracture structures include:

[1] Struktury zhelezorudnykh mestorozhdeniy, (pp.).

Table 1

Types of structures of deposits	Development of structures in time, relative to ore mineralization			Subtypes of structures and examples of deposits
	Pre-ore	Contemporaneous	Post-ore	
I. DEPOSITS OF METAMORPHOGENIC AND COMPLEX GENESIS IN THE PRECAMBRIAN (Krivoy Rog type)				
1. Fold-fracture	Gently sloping regional folds	Complexly folded, plicated, cleavage	Joints	a) Transverse, steeply dipping, open folded structures. Deposits: Artem, Karl Liebknecht, October Revolution (Krivoy Rog Basin) b) Gently plunging crumpled folded structures. Deposits: Dzerzhinskiy (Krivoy Rog Basin), Yakovlev (KMA)
2. Fold-fault	Unclear	Complexly folded, plicated, cleavage, boudinage	Faulted	a) Transverse open folds with fault systems. Deposits: May First (Krivoy Rog), Hunchanlin (Anshan region, Chinese People's Republic) b) Cumpled longitudinal folds with longitudinal and transverse faults. Deposits: Yellow River and Ingulets (Krivoy Rog Basin), certain deposits of the Aldan province, Bakal' group of deposits (Urals)
3. Folded and fold-fault structures of contact deposits	Gently sloping folds, or unknown	Open longitudinal folds	Faults	a) Gently sloping synclinal and anticlinal structures with subordinate development of fault structures. Deposits: Il'icha (Krivoy Rog Basin), Tayga (Aldan province) b) Anticlinal structures cut by transverse faults. Deposits: Tarapak-Likhmanovskiy anticline (Krivoy Rog Basin)

Table 1 (continued)

II. CONTACT-METASOMATIC DEPOSITS (Urals type)				
4. Fold-fault-fracture	Unclear	Open folds, intruded granitoids	Large faults	a) Single anticlinal or synclinal structures in schists-extrusives-carbonate rocks cut by granitoid intrusions. Deposits: Second Northern (Northern Urals), Shilu on Hainan Island (Chinese People's Republic), Mt. Magnitnaya (Urals)
5. Fault-fracture	Gently sloping folds	Faults and fracture zones along contacts of intrusives	Transverse and daigonal faults	a) Major faults with fracture systems on the flanks of large folded structures. Deposits: Vysokogorsk, Lebyazhinskoye, Yestyuninskoye (Urals) b) Zones of crumpling along the borders of various rocks — schists and limestones, extrusives and schists. Deposits: Gornaya Shoriya, Tashelginskaya group, Daye ore field (Chinese People's Republic)
III. HYDROTHERMAL DEPOSITS (Angara-Ilim type)				
6. Fault-fracture	Large regional faults, often filled with basin intrusives	Rejuvenation of faults, formation of fractures. New fault-fracture structures	Rare transverse or diagonal faults	Large-scale regional faults along which basaltic magma was intruded; later veins and stocks of magnetite ores were formed along the fractures in the basic rocks and more rarely in sedimentary rocks. Deposits: Rudnaya Gora, Krasnoyarsk
IV. TRUE MAGMATIC (EARLY MAGMATIC) DEPOSITS (Kusinskiy type)				
7. Faulted or fractured	Unknown	Faulted structures arising in the terminal stage of formation of gabbroic intrusive bodies	Faults	Usually distinct fractures, filled with solid titano-magnetite ores in basic rocks. Disseminated ores in parallel fracture zones

a) Transverse, steeply plunging open folded structures, which are very extensively developed in the Krivoy Rog Basin and are typical of the great majority of deposits of the Saksagan ore district. These are generally major transverse bends embracing two or three, and sometimes four, iron-bearing levels in succession. Such bends may extend along the strike of the rocks for 300 to 500 m, and more rarely for 1000 to 1500 m. The amplitude of the fold is 100 to 200 m, and less often 300 to 400 m. The shape is most frequently simple and open, and more rarely flexural (Figure 1).

At the center of such a bend in the rocks lies the main ore occurrence, composing about 60% of all the ore area of the deposit; on the sides are small associated occurrences that add to the main one. The associated ore occurrences usually have complexly folded structures within the relatively undisturbed and flat, parallel occurrence of the surrounding ferruginous rocks.

b) Gently plunging crumpled synclinal structures frequently containing very large iron-ore occurrences at their hinges. Clear examples of deposits associated with a gently plunging crumpled syncline are the Dzerzhinskiy deposit in the Krivoy Rog Basin, the Galeshchinskoye deposit in the Kremenchug area and the Yakovlev deposit in the KMA, and many others (Figure 2). The area of the closure of the fold some 500 to 700 m up from the hinge is composed of intensively mineralized rocks forming a large ore occurrence. Higher up along the flanks of the syncline, the mineralization of the rocks decreases in intensity, and only within the transverse bends are there ore columns, which dip steeply with the hinge of the fold. Fault structures are found very rarely in these deposits; along them the ore bodies, or small parts of them, are displaced for inconsiderable distances. Sometimes, but not often, there are post-ore thrusts or normal faults. A good example of the association of iron-ore occurrence with a gently-lunging synclinal structure is the Saksagan area of the Krivoy Rog Basin (Figure 3). Here, over an extent of about 30 km, deep shafts (down to 2000 m) and mine workings (down to 800 m) have revealed a crumpled synclinal fold whose core, plunging northward at an average angle of 18° to 20°, is intensively mineralized. From the hinge of this ore occurrence there are periodically spaced branches consisting of groups of ore columns located in areas of transverse deformation — folding of the type shown in Figure 1.

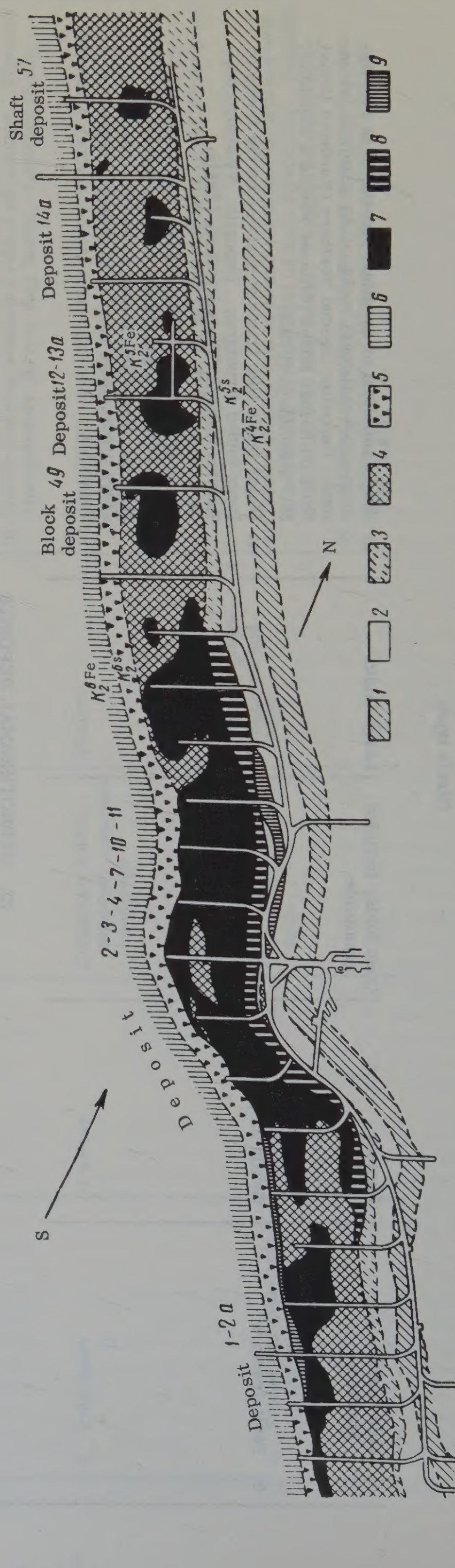

FIGURE 1. Ore deposits located along a flexural transverse bend (Kirov mine, Krivoy Rog Basin).

1 - chlorite-magnetite hornstones of the fourth iron level; 2 - various schists; 3 - goethite-hematite hornstones; 4 - jaspilites of the fifth level; 5 - oreless hornstones of the sixth level; 6 - martite hornstones of the sixth iron level; 7 - martite ores; 8 -

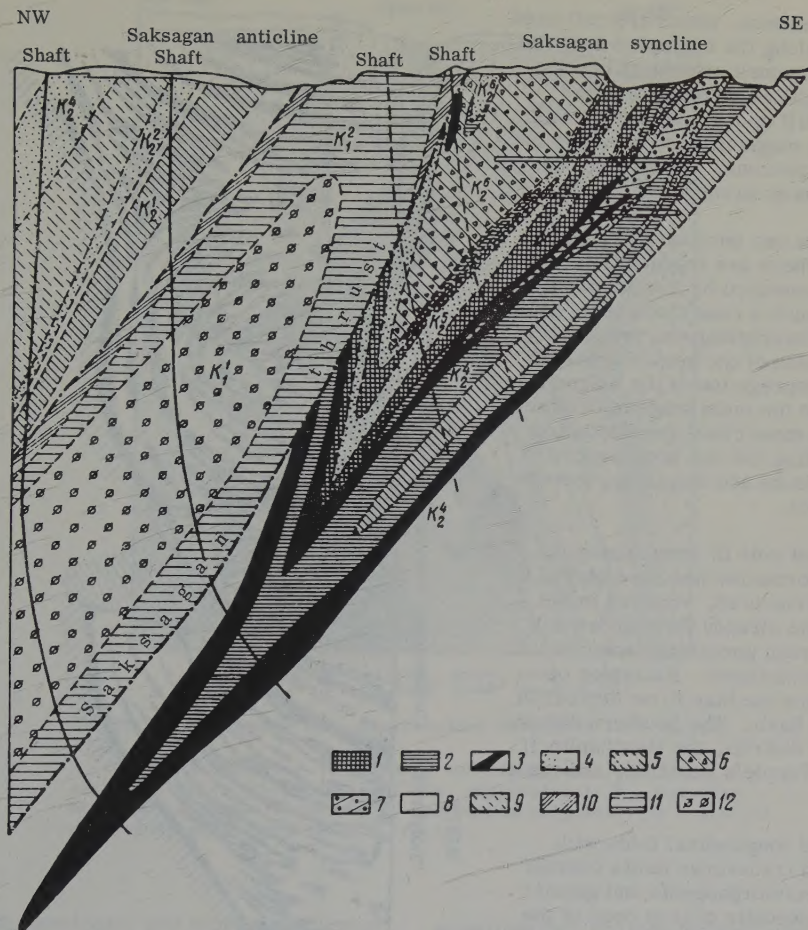

FIGURE 2. Geologic section through the Dzerzhinskiy deposit in the Krivoy Rog Basin.

1 - martite ores; 2 - goethite-hematite-martite ores; 3 - goethite-hematite ores; 4 - chlorite-magnetite hornstones; 5 - oreless hornstones of the sixth level; 6 - chlorite schists; 7 & 8 - various schists; 9 - talc-carbonate level; 10 - phyllites; 11 - arkosic sandstones.

2. Fold-fault structures are found quite requently in ore districts of metamorphogenic rigin, and are also typical of many other kinds f iron-ore deposits. These have various sizes, hapes, positions and degrees of complexity of ne folded structure, and are to one degree or nother broken by faults. The folded structures n such deposits are contemporaneous with the re formation, and are accompanied by thin leavage, intraformational fractures and oudinage, creating the optimal porosity in the ock that will facilitate the circulation of ore-earing solutions. The fault structures in the reat majority of cases are post-ore, displace arts of the ore bodies and produce a very comlicated picture of their emplacement. There re smaller numbers of pre-ore faults, which re renewed during the ore formation and the ost-ore dislocations.

a) Transverse open folds with a system of faults have a complexly folded structure which is generally reflected as a large cross fold within which all the ore occurrences of the deposit are located. This large fold is complicated by smaller bends, as well as by plications and gaps. The ore occurs in the core and on the flanks of the main syncline, in places where the rocks have been subjected to additional bending and form gently sloping open structures and flexures. In addition, the structure of such deposits is usually embellished with numerous longitudinal, cross and diagonal faults. Often the latter have branching forms. Branching out from the main fault are oblique smaller faults, which in turn branch out into still smaller fractures (Figure 4).

In some cases it is possible to observe

pre-ore displacements, which are reflected in the fact that along the crumpled zone the ferruginous hornstones or schists contain abundant segregations of magnetite or hematite which fill up the fractures. The formation of the magnetite and hematite is an ore-forming process, which produced the ore occurrences arranged in a row.

In other cases one encounters crumpled zones in which there are fragments of hematite ores cemented by fine-grained magnetite. In such a case there is no doubt that the zone of crumpling was produced after the formation of the hematite ore, but before the segregation of the magnetite, which belongs to the main stage of ore mineralization. In most cases the faults are post-ore, breaking the ore occurrences into separate blocks and displacing them to various distances.

Thus the chief role in determining the regime of ore formation has been played by fold-fracture structures, whereas in the disposition of the already formed ore occurrences the main governing factor has been the fault structures. Examples of such deposits are the May First deposit in the Krivoy Rog Basin, The Southern deposit in the Belozero district, the Hunchanlin-II in the Chinese People's Republic, and many others.

b) Crumpled longitudinal folds with longitudinal and transverse faults contain many of the metamorphogenic and genetically complex deposits of iron ores of the Precambrian. Usually these are vertical or steeply inclined folds in ferruginous clay rocks, complicated by transverse or longitudinal faults. They include complex flexural folds, large synclines or systems of complex and sometimes fan-shaped folds complicated by numerous faults (Figure 5). The latter are represented by thick longitudinal zones of crumpling and gentle or steep transverse shear fractures, along which there have been clear displacements of the rocks and ore bodies. The iron ores occur primarily in the synclines, in the form of complex metasomatic bodies that extend to great depths as ore columns. The borders of the ore bodies repeat the complex folds in the rocks, and they are usually separated along the flanks of the folds into apophyses or satellite ore bodies. Structures of this subtype include the following deposits: the Yellow River and the Inguletskoye in the Krivoy Rog Basin, some of the deposits in the Aldan province, and also the Bakal' group of siderite deposits in the Southern Urals.

3. <u>Folded and fold-fault structures of contact zone deposits.</u> These are deposits

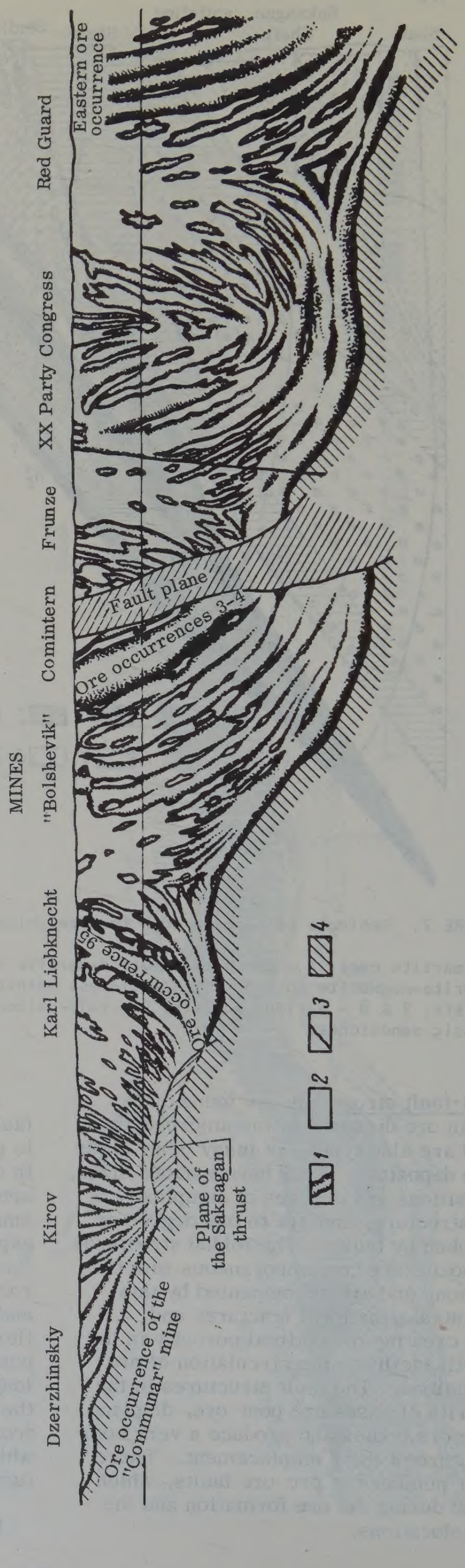

FIGURE 3. Diagram of the distribution of ore occurrences in the Saksagan syncline.
1 - iron ores; 2 - ferruginous rocks; 3 - diabase dikes; 4 - chlorite-sericite schists.

FIGURE 4. An example of a fault-folded structure in a deposit.

1 - magnetite ore; 2 - ferruginous hornstones and jaspilites; 3 - various schists (chlorite, sericite, amphibolite); 4 - breccias and zones of mylonitization; 5 - carbonate-talc level; 6 - arkosic sandstones and quartzites; 7 - faults.

:ith bed-like, lenticular and more complicated re bodies, usually localized at the boundaries f the iron-ore rocks that overlie schists or enesis. The distinctive feature of these de-osits — their sedimentary-metamorphic enesis and their simple monoclinal structure - is in many cases complicated by folds and isjunctive dislocations. Moreover the ore odies are not found to be located primarily in ynclines, anticlines or areas of intensive fault-ng. The formation of the ore bodies in these eposits and their composition is directly con-ected with the facies, paleogeography and the ubsequent metamorphism of the region.

The following subtypes exist among the struc-ures of these deposits:

a) Gentle synclinal and anticlinal structures 'ith a subordinate development of faults are ypical of numerous deposits in the Southern ore eld of the Krivoy Rog Basin along the Eastern igulets syncline, and of the Tayga deposit in ie Aldan province (Figure 6).

The ore bodies of these deposits are mor-hologically represented by beds, lenses and iore rarely by thick ore columns within the stratified ore-bearing formation or level. The general structure of the deposit is a syncline, often with gentle second-order bends. Faults break up the ore bodies into blocks of various sizes and displace them relative to each other, in places thus forming a very complicated block structure. In the overwhelming majority of cases the folded structure plunges gently and retains its second-order folds for great distances. The intensity of the mineralization in the producing level is constant, and changes or comes to an end only with a change in the lithologic composition and the nature of the folded structure.

b) Anticlinal structures broken into blocks by transverse faults. Deposits of this subtype are located in the core or on the flanks of the anticline, and are confined, as in the previous case, to the producing level or formation. The steep anticline is divided by many transverse dislocations into blocks. Each block is in turn broken up into a multitude of small chunks that are displaced by various distances (Figure 7). The northern blocks are displaced toward the acute angle formed by the bedding planes and the plane of the dislocation. The faults are marked by fractures of various sizes, filled

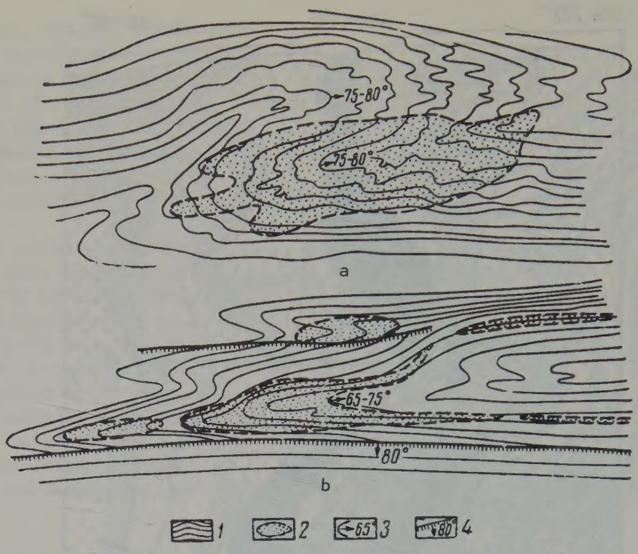

FIGURE 5. Diagram showing the disposition of magnetite ore occurrences within compression folds.

a - flexure; b - syncline with faults; 1 - bedding; 2 - iron ore occurrences; 3 - buried hinges of folds; 4 - dips of faults.

with vein quartz, carbonates, silicified blue asbestos and magnetite.

II. CONTACT-METASOMATIC DEPOSITS (URALS TYPE)

The structures of contact-metasomatic bodies are determined by the tectonic and magmatic conditions under which they have originated. The formation of deposits of this type is closely associated with the injection of moderately acidic magmas into sedimentary or sedimentary-extrusive rocks rich in calcium, which in melting lowered the solubility of the metals and withdrew from the melt the iron that existed there in ionic form. At this stage the

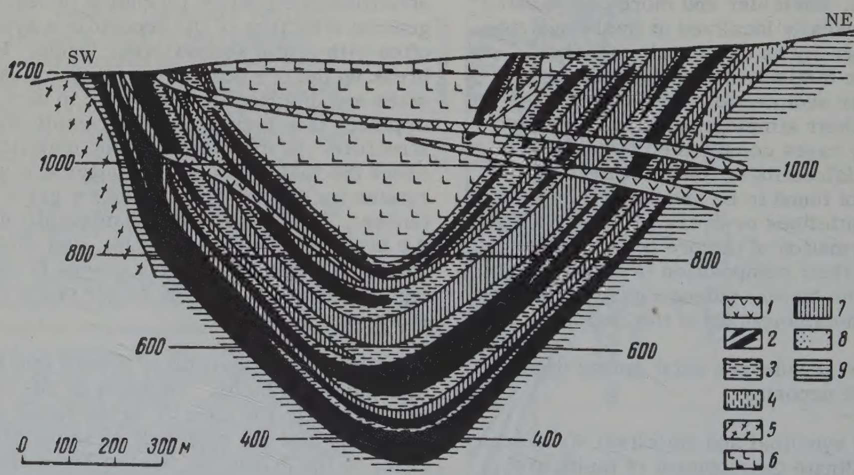

FIGURE 6. Diagrammatic section through the Tayga deposit along the line A - B (after V.A. Pervago).

1 - syenite porphyry; 2 - magnetite ores; 3 - magnesium skarns; 4 - granitized rocks; 5 - migmatites with injected granites; 6 - sillimanite gneisses of the supra-ore level; 7 - biotite and amphibole gneisses, partially altered to skarn; 8 - calcipyres and dolomitized marbles; 9 - schists and pyroxene-amphibole gneisses.

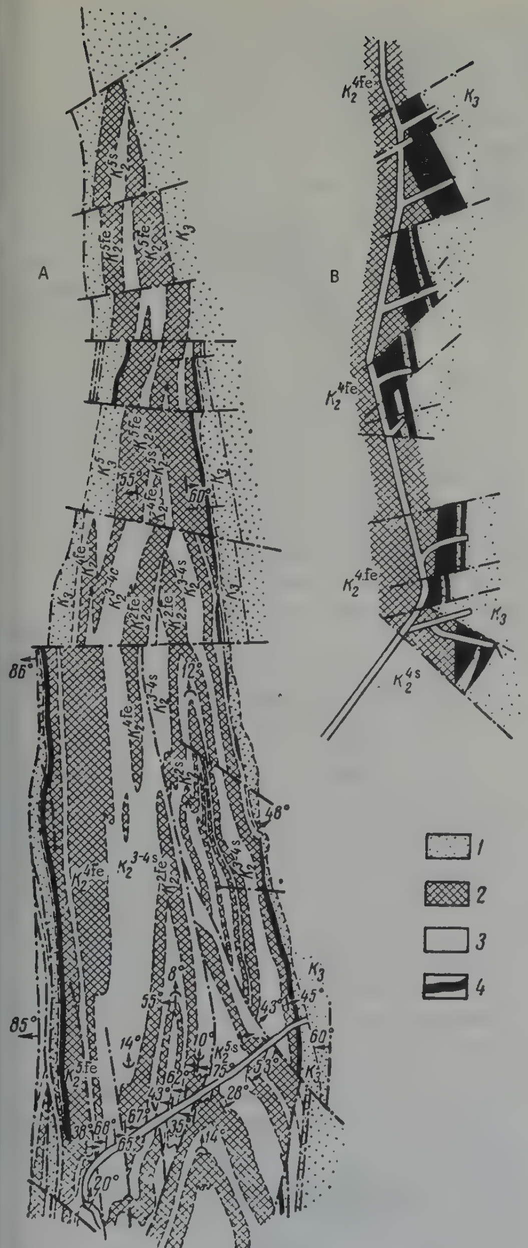

FIGURE 7. Fold-and-fault structure of the Tarapak-Likhmanovskiy anticline (A) and a detailed diagram showing the development of faults, according to data derived from mining operations (B).

1 - carbonaceous and sericite schists; 2 - chlorite-sericite and amphibole schists; 3 - ferruginous hornstones and jaspilites; 4 - magnetite ores.

rocks began to heat up, thus increasing their plasticity and limiting the penetration of solutions into the rock. Cooling of the rocks then led to decrease in their plasticity, accompanied by intensive jointing and sharply increasing the rocks' permeability to solutions.

With the development of the tectonic movements, the rock becomes increasingly permeable both along newly formed fractures and along the bedding planes and previously existing pores. In many deposits of Siberia, the Urals, China and other countries the ore occurrences are regularly located in areas of rock altered by contact phenomena.

The areas favorable to the localization of ore bodies are the following:

a) The boundaries between skarns and unaltered rocks or skarns and intrusive rocks. Here the largest ore bodies are located. The ore occurrences within he skarns are usually small and poor in iron, thus belonging to the secondary or low-grade category. This is most likely to be explained by the fact that in the high-temperature stage of skarn formation the iron taken from the melt together with the rock-forming elements in the form of Fe^{2-} and Fe^{3-} has gone into the formation of silicate minerals. In the later stage, as a result of the decrease in the solubility of the metals, the iron that occurs in ionic form or, perhaps, in compounds with metallic bonds, is segregated as tiny droplets and removed by gas bubbles, as in the process of froth flotation [6]. The gas and liquid products are removed in the form of streams along the most favorable tectonic paths to areas of lower pressure, where the iron-ore minerals can be deposited to form large ore bodies. These tectonic pathways are the zones of fractures that arise during this process at the boundary between the intrusive and the skarn or between the skarn and the unaltered surrounding rocks, caused by the considerable differences in their physical properties.

b) Steeply dipping contacts of limestones or steep projections of the limestones into the intrusive rocks. These are the areas most highly shattered tectonically, broken by numerous fractures oriented along the contact [1].

c) Contacts between limestone or marble strata and the underlying extrusives, intrusives or schistose sedimentary rocks. In such places are usually concentrated the large stratified or bed-like ore occurrences that are associated with systems of bedding-plane fractures.

d) Limestone interbeds in strata of tuff-sediments whose composition is favorable to ore mineralization; this is also facilitated by the appearance of a multitude of shear and tension fractures.

e) Steeply and gently dipping (relative to the folded structure) pre-ore faults that are often filled with crushed and shattered skarn minerals cemented by magnetite or sulfides.

Papers in recent years produced by many staffs of mining geologists and research institutions have shown that the great majority of industrial deposits of this type are located in the cores of anticlines and synclines or in zones of faulting and crumpling developed on the flanks of major folds.

Thus contact-metasomatic deposits of iron ores arise under a favorable combination of conditions, including moderately acidic intrusions into calcium-containing rocks with fold-fracture or fault-fracture structures that make the rocks permeable to ore-bearing solutions and favor the course of metasomatic processes.

Almost all the contact-metasomatic deposits are characterized by fold-fault-fracture and fault-fracture types of structures.

1. Fold-fault-fracture structures in iron deposits occur quite frequently, and are characterized by single anticlinal or synclinal folds composed of carbonate or other calcium-containing rocks. These folded structures are cut directly by the intrusives or lie in close proximity to them. Many ore districts are characterized by the intrusion of granitoids into the cores of the anticlinal or brachysynclinal folds, with branches of these intrusives extending into the flanks. The ore-bearing channels were, in all cases, interformational and intraformational fractures and steep faults that appeared at the concluding stage of mineralization. The ore was deposited under favorable conditions in those parts of the level where there were enough small fractures, pores and capillary micropores through which the solutions could be dispersed to react with the rock. After the ore bodies were formed, the former faults were renewed and new ones appeared, resulting in the creation of the very complex fold-fracture-fault structure of the deposit.

The most typical representatives of the structures of this type are the deposits of Mt. Magnitnaya, the First and Second Northern mines in the Urals, Shilu on Hainan Island (Chinese People's Republic), Nurdmark in Northern Sweden, and many others. The Mt. Magnitnaya deposit is associated with an anticlinal uplift that is complicated by numerous normal faults. Submeridional and subequatorial faults break up the ore body and give it its block-like character. Along the eastern slope of Mt. Atach a large meriodional normal fault with an amplitude of about 250 m divides the eastern part of the deposit from the western. There is a series of small parallel faults with amplitudes of 10 to 30 m which give the flanks of the anticline their step-like shapes. Keratophyre and diorite dikes are developed in the submeridional faults.

According to many indications, the submeridional faults were formed before the ore and some of them were renewed afterwards; this is supported by the following facts: a) the dikes of diorite within the submeridional faults are in some cases replaced by magnetite, and b) in many places the submeridional faults are filled with ore material. On the other hand, the subequatorial faults are post-ore, open or ground down, and the displacements along them are of small amplitudes.

The deposit of the Second Northern mine is represented by magnetite ore bodies, located in the crests and trough of folds, where they have replaced lenses of fractured marble limestone (Figure 8). The Shilu deposit occurs in a stratum of Permian limestones forming a large synclinal and anticlinal fold near the contact with the Yanshan granites (Figure 9). All the ore bodies of this deposit are located in the limestone stratum, regardless of the contact with the granites. The thickest ore occurrences are in the core of the anticline; the thinner ones are in its flanks and are completely absent from the core. The fault near the deposit probably also served as an ore-conducting structure.

2. Fault-fracture structures are also very typical of contact-metasomatic iron-ore deposits. They are formed in the terminal and concluding stages of the tectonic process in the formation of mobile zones. Their peculiarity consists in the fact that the large ore-bearing faults or zones of crumpling with systems of accompanying fractures arise near the intrusive contact of the granites, syenites or diorites on the flanks of large folds or in areas where the rocks have a monoclinal structure. In certain areas there are tension fractures in the apical part of the intrusive, in the direction of its long axis, in the form of radially diverging rays. The faults or zones of crumpling that were produced before or at the same time as the formation of the ore continue to be active during the main period of ore mineralization, and in many cases are renewed after the formation of the deposit.

Quite often it is difficult to judge the presence or absence of ore-controlling pre-ore zone and to map it. In many cases, however, from a number of indications one can nevertheless reliably speak of the pre-ore age of certain zones of crumpling or faulting in the deposit, especially on the basis of data from mine workings or open-pit operations. These indications include: a) segregations of magnetite or other ore and associated minerals within the zone, replacing the material or filling fractures; b) lack of correspondence of the contours of a section through an ore body located on both sides of a fault zone, in contrast to post-ore faults, in which case the contours of the section through

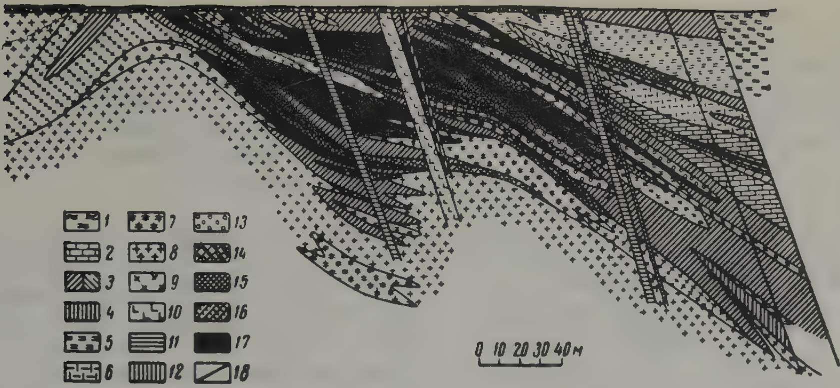

FIGURE 8. Geologic-lithologic section through the Vtoroy Severnyy mine (after Ya.P. Baklayev).

1 - weathered crust of extrusive rocks; 2 - limestones; 3 - porphyrites; 4 - porphyries; 5 - porphyrite tuffs; 6 - tuffs breccias of extrusive rocks; 7 - extrusive rocks; 8 - granodiorites; 9 - diorites; 10 - granodiorite porphyries; 11 - diabases; 12 - diabase porphyrites; 13 - skarns; 14 rocks containing disseminated magnetite (Fe = 20%); 15 - disseminated magnetite ores (Fe = 20-30%); 16 - magnetite ores (Fe - 30-45%); 17 - massive ores (Fe - 45%); 18 - lines marking tectonic dislocations.

the ore body on both sides of the fault must coincide; c) a linear elongation of the ore bodies and the products of mineralization during the ore period in the direction of the zone of dislocation; d) veins and dikes of diorites, and porphyrites striking in the same direction as the ore bodies; e) replacement of the selvage of veins or dikes by ore mineralization; f) a general orientation of the minerals in the same direction as the elongation and disposition of the ore bodies.

The finding and tracing of the ore-controlling structures within the deposit and beyond it is of great importance in expanding the known resources of ores. If during the time of the ore mineralization the faults and fractures created the conditions required for the accumulation and concentration of ore material, then in post-ore times the ore occurrence will frequently be broken into numerous blocks along new or renewed earlier faults; in some cases this will hinder the exploitation of the deposit and in other cases even make the deposit economically worthless.

Two subtypes of deposits associated with fault-fracture structures are recognized:

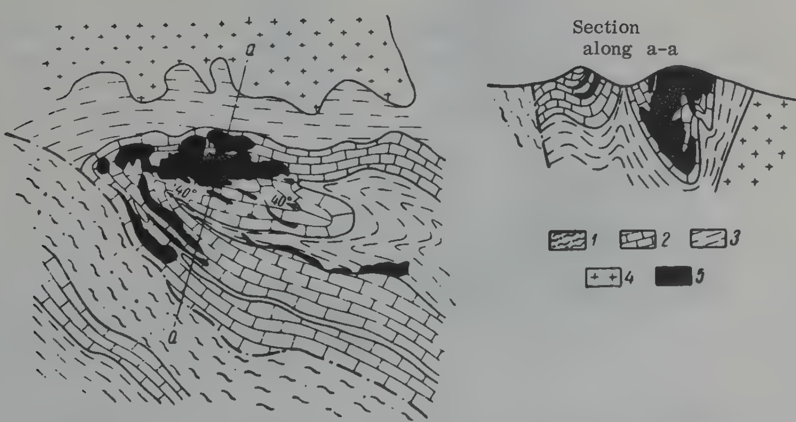

FIGURE 9. Diagram of the geologic structure of the Shilu deposit on Hainan Island, Chinese People's Republic.

1 - schists; 2 - limestones; 3 - quartz-sericite schists; 4 - Yanshan granites; 5 - ore.

a) Large faults with fracture systems, developed usually on the flanks of major folds, are distinguished by the presence of a fault or fault system, with a net of fractures diverging from it, earlier or contemporaneous with the ore. The area of this fault and its fractures is the place in which the iron-ore deposits of this type are emplaced. The fracture system is filled in; around it there are traces of infiltrational and diffusional metasomatism in the calcium-containing rocks. This subtype is often characterized by the development of numerous post-ore displacements creating a complex block structure. Representatives of this subtype are, among others, the deposits of the Tagil-Kuvskinskiy group in the Middle Urals, among which the Goroblagodatskoye deposit has a large pre-ore fault between Silurian limestones and tuffs. This contains a belt of reworked limestones and tuffs that have been transformed into skarns and iron ores. The strike of the ore occurrences follows that of the main open-pit working, west and east of the Revdyansk area. They are broken by numerous post-ore displacements that create a complicated mosaic of blocks in the deposit (Figure 10). It should be noted that here the pre-ore zones are very difficult to recognize, since they are filled with the ore and the accompanying mineralization.

b) Zones of crumpling, sometimes with post-ore faults, are widespread in many contact-metasomatic iron-ore deposits. Most frequently at the boundary between rocks of different composition and physical properties there is a broad zone of crumpling, manifested as a system of numerous shear and tension fractures. Such zones also often occur at the boundary between crystalline limestones and schists, marbles and intrusive bodies, limestones and tuffs, and many other rocks. Cases are not rare in which the crushed zone passes directly into the granite or

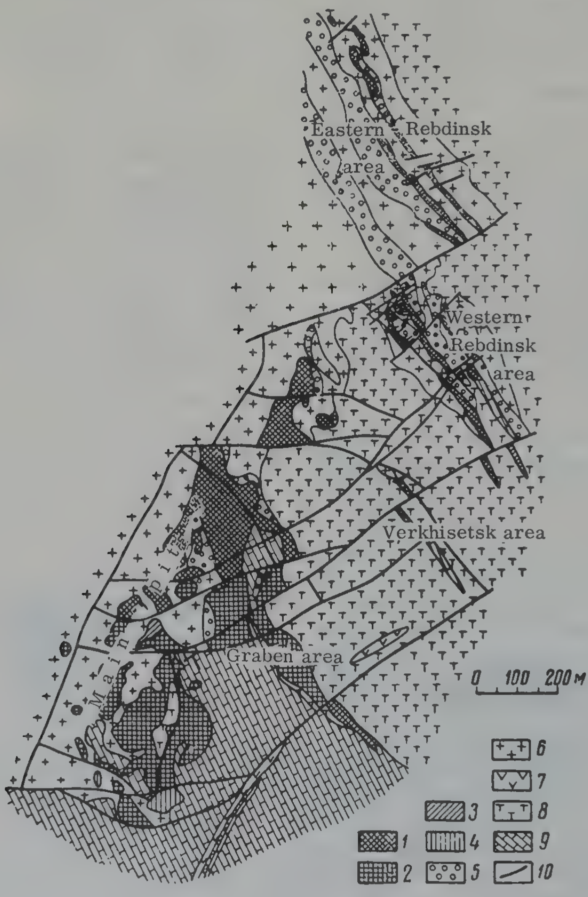

FIGURE 10. Schematic tectonic map of the Vysokogorsk iron-ore deposit (after L.N. Ovchinnikov).

1 - magnetite ores; 2 - martite ores; 3 - hemimartite ores; 4 - limonites; 5 - oreless skarns; 6 - syenites; 7 - porphyrites; 8 - tuffs and tuff breccias; 9 - limestones; 10 - lines marking post-ore dislocations.

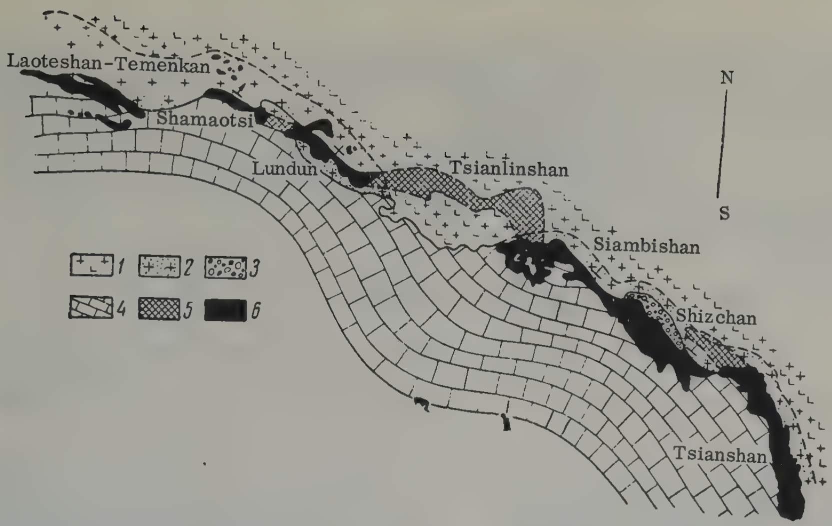

FIGURE 11. Diagram of the emplacement of the ore bodies in the Dae-Teshan iron deposit, Chinese People's Republic.

1 - diorites; 2 - altered diorites; 3 - skarns; 4 - marbles; 5 - blind (hidden) ore occurrences; 6 - iron-ore bodies.

enite massif, commonly in places where the of of the intrusive has preserved xenoliths. contrast to the preceding subtype, deposits sociated with crushed zones rarely contain st-ore faults. This is likely due to the fact t the crushed zones were themselves formed the latest movements, which occurred in the ry latest stages of formation of the folded nes.

Deposits belonging to this subtype include ose of the Tashelginskaya, Mayzasskaya and rnaya Shoriya groups (Sheregesh, Shalym, shtagol, Kochura), and also the deposits of Daye ore field in Central China (Figure 11) d many others. The Daye-Teshan deposits veal a very interesting phenomenon: the ore dies occur in those parts of the marbles that oject into the diorites.

III. HYDROTHERMAL DEPOSITS (ANGARA-ILIM TYPE)

It was pointed out above that hydrothermal posits amount to a very small part of the total sources of iron ores. Nevertheless in the S. S. R. and a number of other countries a rtain number of industrial iron-ore deposits e hydrothermal; this contention is, however, sputed by some geologists.

All the deposits or iron ores of this genetic oe are characterized by fault-fracture structes. These are often very large tectonic faults accompanied by pre-ore intrusions of basic rocks, or else a system of normal faults in sedimentary rocks. The period of mineralization coincided with the reactivation of the old structures and the formation of new faults with systems of fractures diverging from them. The post-ore structures take the form of transverse or diagonal faults. The structures of this type are clearly represented in the deposits of the Siberian platform.

Within the Tungusska synclise have been found several iron-ore districts, located mainly at its western and southwestern margins. This is the location of the Tungusska iron-ore field, the ore district of the Bakhta and Podkamennaya Tungusska Rivers, the Angara-Ilim district and the Ilimzhi ore district [8]. All these ore districts are associated with zones of regional faulting fracturing and intensive igneous activity along the borders of the synclise. The structures of the ore districts consist of flat-lying Paleozoic sedimentary rocks and tuffs with isolated narrow anticlinal folds. These folds are intensively broken up by steeply dipping disjunctive dislocations — fractures, faults and crushed zones. The magnetite ore bodies occur in the zones of weakness, their strikes and dips being the same as those of the tectonic zones along which the magma passed before the ore mineralization; this magma formed amphibolitized gabbro-diabases. Magnetite ore bodies in the form of veins, lenticular accumulations and stockwork bodies of various sizes intersect both the traprocks and the Paleozoic sedimentary rocks that enclose the volcanic pipes.

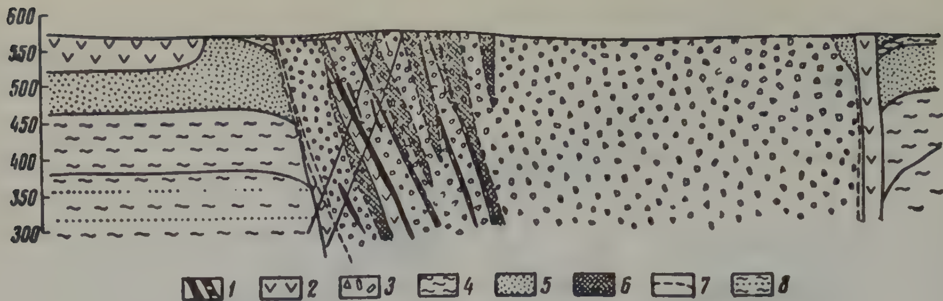

FIGURE 12. Krasnoyarsk deposit (Angara-Ilim region), geologic section (after N.V. Pavlov).

1 - magnetite ore veins; 2 - gabbro-diabases; 3 - tuffs and tuff breccias; 4 - argillites; 5 - sandstones; 6 - disseminated ores; 7 - pre-ore faults; 8 - sand-clay rocks.

It has been observed that the mineralization is localized in zones of disjunctive dislocation that are filled with differentiated traprock, and also in the accompanying weakened zones that frequently follow the bedding of the sedimentary rocks; the magma and the hydrothermal solutions penetrated through these weakened zones.

In the Rudnogorsk, Korshunov and other deposits considerable masses of ore are enclosed in rocks that fill the pipe-shaped volcanic formations. Geologists who have studied these deposits believe that the pipe-like bodies filled with pyroclastic material go far down into the depths and that later tectonic movements in them created zones of weakness and cavities in them that were especially favorable to the passage of the ore-bearing solutions.

Thus the general characteristic feature of the iron-ore deposits of the Tungusska synclise is the close connection between the mineralization and the preceding disjunctive dislocations. The deposits were formed by ore-bearing solutions emerging from basaltic magma chambers at depth. The paths along which the solutions passed were tectonically renewed faults along which the basaltic magma penetrated before the formation of the deposits. In places the ore mineralization was localized both in fault zones and the accompanying zones of weakness paralle to the bedding, and in the pipe-like bodies (neck occurring in the more mobile parts of the Tungusska synclise (Figure 12). Chemically th Angara-Ilim deposits differ essentially from the vein deposits associated with a granitic magma in that the silicate plays a very subordinate role in the former, and the elements potassium, sodium, lithium and boron are completely abser Moreover the ores of the Angara-Ilim district are characterized by the presence of titanium, vanadium and cobalt, in addition to the usual iron, calcium, magnesium, water, carbon dioxide gas and phosphorus (chlorine and fluorine)

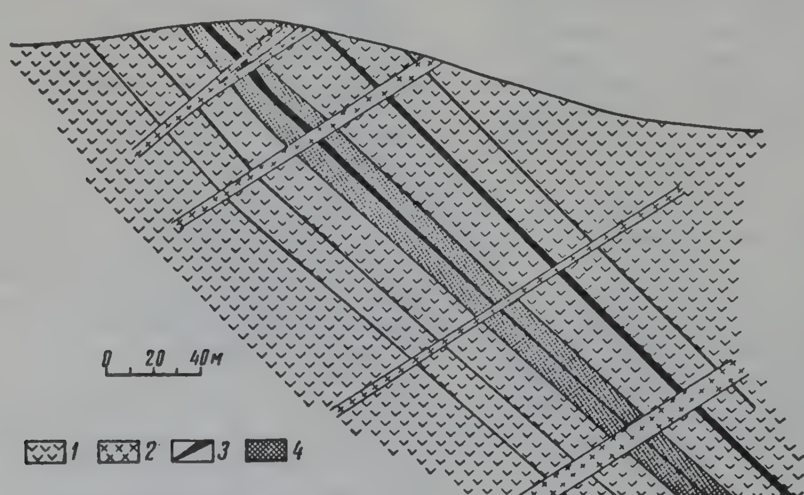

FIGURE 13. Geologic-lithologic section through the Kopan deposit.

1 - gabbro; 2 - diabases; 3 - rich ores; 4 - poor disseminated ores.

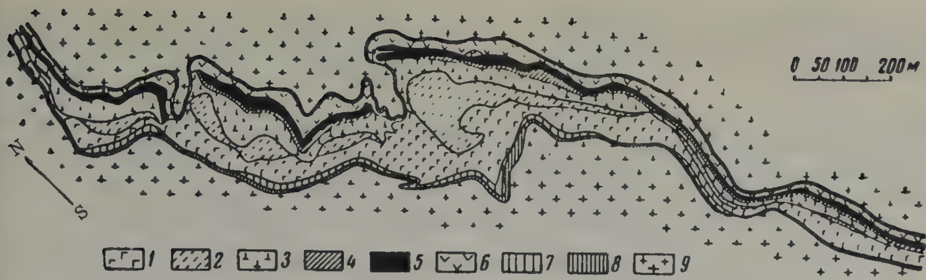

FIGURE 14. Lithologic-geologic map of the Pudozhgorsk titanomagnetite deposit.

1 - metabasites; 2 - uralite level; 3 - fine-grained metadiabases; 4 - metagabbro-diabases; 5 - titanomagnetite ores; 6 - diabases; 7 & 8 - aphonite metadiabases; 9 - plagioclase-microcline granites.

7. TRUE MAGMATIC (EARLY MAGMATIC) DEPOSITS (KUSINSKIY TYPE)

The structures of this category in all or al- ost all the titanomagnetite deposits are char- terized by single or series of fault fractures at are morphologically distinct. These frac- res, located in gabbroic massifs or meta- sites, are filled with solid titanomagnetite es (Figure 13). In the Kusinskiy deposit such acture-filling ore bodies extend for several ometers as three parallel veins. Downward ey wedge out or are replaced by parallel frac- res that do not reach the surface.

In the Pudozhgorsk deposit the veins of anomagnetite ores are located in a system of foliation fractures in the contact zone (Fig- e 14). In addition, deposits of this type con- n bodies of low-grade disseminated ores rep- esenting zones of very small fractures; some vestigators believe that the ore material was cumulated in these small fractures when it s squeezed out from the very small spaces in e last stages of formation of the gabbroic assif.

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SOME REGULAR ASPECTS OF THE FORMATION OF THE KARKARALINSK INTRUSIVE COMPLEX[1]

by

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On the basis of a geological and petrochemical study of the Karkaralinsk intrusive complex, it as been established that there is a regular successive change in the chemical compositions of the ntrusives; the contact relationships of the intrusive masses of various ages, accompanied by processes of assimilation and replacement of the exocontact rocks, are also considered.

* * * * *

A. PRELIMINARY REMARKS

In the eastern part of Central Kazakhstan, in he Karkaralinsk district of the Karaganda egion, lie the Karkaraly Mountains. These nountains and their foothills are composed rimarily of intrusive rocks, which penetrate edimentary deposits of the Paleozoic; in addition to these there are some extrusives of diferent composition.

The study of the Karkaralinsk intrusive complex was accompanied by geologic mapping, which formed the basis for the preparation of a etrographic sketch map of the complex.

Chemical and semiquantitative spectrum nalyses were made in the Laboratory of the nstitute for the Geology of Ore Deposits, etrography, Mineralogy and Geochemistry of he Academy of Sciences of the U. S. S. R. The ata resulting from the chemical analyses of he igneous rocks were computed by the A. N. avaritskiy method. Quantitative mineral ounts on the thin sections were made by O. S. ogomolova.

Very few papers on the subject of the Karainsk intrusive complex have been published in he literature of geology.

At the end of the nineteenth and beginning of he twentieth centuries N. K. Vysotskiy [4] and . A. Anosov (1916) made a regional geologic nvestigation of the Karkaralinsk area. N. K. azumovskiy [9] later visited the northwestern art of the Karkaraliny Mountains. An artical y M. P. Rusakov [10] includes a geologic suvey map on the scale of 1:840, 000, which shows the intrusion of the Karkaralinsk granites. In 1936 N. I. Nakovnik [6, 7] discovered some occurrences of tungsten ore in the endocontact greisens of the granitic massif north of Mt. Karkaralinsk.

The results of later investigations have been fairly well summarized in an unpublished paper by V. I. Yagovkin (1955). About 1. 5 km northeast of Mt. Karkaralinsk he collected a brachiopod fauna which M. A. Borisyak and N. L. Bublichenko have identified as typical of the Lower Devonian. In the northwestern exocontact of the Karkaralinsk intrusive complex, A. G. Timofeyev discovered a fauna belonging to the lower strata of the Upper Devonian in some limestone interbeds. A fauna of presumably Upper Devonian age has also been found by V. I. Yagovkin on the left bank of the Karkaralinka River, east of the Karkaralinsk intrusive massif. V. I. Yagovkin believes that the Devonian deposits form the lower structural stage of the Paleozoic strata, which are unconformably overlain by younger, Middle Carboniferous, extrusives.

Among the intrusive rocks V. I. Yagovkin distinguishes granodiorites and quartz diorites, dike intrusives of syenite porphyries and granosyenite porphyries, and Late Hercynian alaskite granites.

B. STRATIGRAPHY

1. Middle Paleozoic. The sedimentary deposits northeast of Mt. Karkaralinsk, as already mentioned, have been identified as Devonian, on the basis of their content of fauna belonging to the Lower Devonian. Lithologically this series is characterized by the predominance of black siltstones interlayered with gray fine-grained

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[1]Nekotoryye zakonomernosti formirovaniya Karkaralinskogo intruzivnogo kompleksa, (pp. 21-38).

and medium-grained sandstones and dark clay shales; it contains not only Lower, but also in part Middle Devonian fauna. Similar deposits also occur some 8 to 10 km south of Mt. Karkaralinsk, in the form of a zone trending subparallel to the equator and extending from Mt. Nayzy to the Komsomolets kolkhoz (Figure 1).

FIGURE 1. Petrographic sketch map of the Karkaralinsk intrusive complex.

Prepared by N.F. Anikeyeva, A.T. Vasilevskiy, Ts.Van. N.A. Voronina, V. Den, L.Ye. Lynov, Ye.A. Maksimov, under the direction of N.F. Anikeyeva, 1955-1958.
1 - Quarternary deposits; 2 - alaskite and aplitoid granites; 3 - biotite granites intruded by alaskite granites; 4 - biotite granites, hybrid granosyenite porphyries and granosyenites, quartz syenite porphyries and quartz syenites; 5 - rhyolite-trachte porphyries and andesite-basalt porphyrites; tuff breccias; 6 - quartz syenite-diorites, quartz diorites and granodiorites; 7 - conglomerates, sandstones and shales; 8 - metamorphosed sandstones and shales with interlayered rhyolite porphyries, diabase porphyrites, tuffites and skarn limestones; 9 - metamorphosed sandstones and shales; 10 - zones of disjunctive tectonic dislocations.

The upper strata of the sedimentary series contain lilac-gray and greenish-gray rhyolite porphyries, accompanied by light-colored psammitic and pelitic tuffites forming a layer of varying thickness among the sandstones and shales. This same series sometimes contains greenish-gray porphyrites of the spilitic type, tuffs and tuffites of basic composition, as well as interlayers of skarn limestones whose thickness ranges from a few meters to a few tens of meters. By analogy with the adjacent areas, this series with its interlayers of extrusives and limestones may be tentatively called Middle to Upper Devonian.

At the northwestern contact of the Karkaralinsk complex there are Upper Devonian deposits. Here, more extensive than the limestones in which the fauna were discovered, are dense small-pebble conglomerates and sandstones of greenish-gray and varied colors, with interlayers of metamorphosed shales.

2. Upper Paleozoic. In addition to the strata described above, the area also contains light gray rhyolite-trachyte porphyries which form thick flows, interlayered with flows of black andesite-basalt porphyrites which are usually some tens of meters thick. The extrusives are accompanied by tuff breccias of mixed composition.

The ages of these extrusives have been differently interpreted: some geologists assign them to the Devonian, and others (V.I. Yagovkin) to the Middle Carboniferous. The investigations made by the present writer have established a close physical connection between these extrusives and dikes of similar composition, which are sometimes represented by rhyolite-trachyte porphyries and andesite-basalt porphyrites, but are more often completely recrystallized. These dikes are apparently the roots of extrusive outpourings and cut through not only the Devonian strata, but also the Early Hercynian intrusives of quartz syenites-diorites, suggesting that the extrusives are most likely of Late Paleozoic age. They are apparently parallel in age to the Late Paleozoic extrusives of the Balkhash volcanogenic series.

The section through the Paleozoic strata ends with these extrusives, which were poured out under terrestrial conditions.

Apart from the Paleozoic strata, the area under consideration here contains only unconsolidated Quaternary deposits.

C. TECTONICS

Tectonically the area is a part of the northern margin of the Dzhungara-Balkhash geosyncline and lies west of the Chingiz anticlinal upuplift. In the Karkaralinsk region, the northwestern structures of the Chingiz Range are replaced by folding trending roughly east-west.

The Devonian deposits that contain the Karkaralinsk intrusive complex are crumpled into folds that strike along the azimuths NW 290° - SE 70°. The sharp deviations from these azimuths are due to the plunges of the axes of these folds or to disjunctive dislocations.

The northern part of the area is apparently associated with the subequatorial anticline, a considerable part of which was later destroyed by the intrusive processes. The axis of this anticline probably emerged at the meridian of the city of Karkaralinsk, plunging at the western and eastern boundaries of the area, so that to the northeast of Karkaralinsk appears its core, composed of sedimentary rocks with Lower Devonian fauna.

In the southern part of the area lies a second anticlinal fold trending NW 290°; this passes through the vicinity of the Komsomolets kolkhoz, somewhat south of Mt. Nayza. At the meridian of Karkaralinsk the axis of this anticline also emerges, so that to the north of the Kyzylshoky kolkhoz the core of this fold also contains sedimentary rocks analogous to those of the Karkaralinsk fold. Where its axis emerges, the flanks of the anticline are composed of younger rocks, of Middle and Upper Devonian age. The southern flank of the fold clearly dips southward, but the northern flank is complicated by numerous disjunctive dislocations and intrusive granitoids that obscure its mode of occurrence. The axis of the anticline plunges eastward; the sedimentary strata form a periclinal structure and in the vicinity of Mt. Nayza are replaced by younger extrusive-sedimentary deposits of Middle to Upper Devonian age, which trend approximately parallel to the equator. The anticline apparently also plunges westward, where the section through the strata along its axis contains extrusives and skarn limestones (northwest of the Komsomolets kolkhoz).

Between the anticlines is a synclinal region, with is filled primarily with Upper Paleozoic extrusives and penetrated by various intrusives. The Upper Paleozoic extrusives have also been folded, but evidently less intensively than the structural stage underlying them, which is composed of Devonian strata.

The area of the Karkaralinsk intrusive complex is characterized by a great abundance of disjunctive tectonic dislocations on various scales. There is a predominance of northeastward trending tectonic zones, which form a series of extensive parallel faults with constant azimuths of 50 to 60°. The tectonic zones with northeastward trends are regularly combined with perpendicular northwestward zones trending along azimuths of 310 to 320°. In addition

to these zones, there are also disjunctive dislocations with different trends. Some of these are subordinate to the main northeastward and northwestward zones; others are independent.

The tectonic zones were formed in various stages. The intrusions of thick, dike-forming bodies (γ_2) are evidently to a considerable degree associated with the then already existing zones of tectonic dislocations. The intensive injections of alaskite granites (γ_3) in older rocks are apparently also connected with tectonic zones. On the other hand, the tectonic movements did not come to an end with the crystallization of the alaskite granites, since the latter are intersected by tectonic zones with the formation of alaskite breccias and mylonites cemented by the products of their post-magmatic activity. The quartz veins and veinlets that cement the breccias in turn bear traces of later tectonic movements.

D. INTRUSIVES

1. The Succession of Intrusive Phases

The Karkaralinsk intrusive complex was formed in a number of intrusive phases. The earliest phase comprises the intrusion of the quartz syenite-diorites, into which were intruded the later biotite granites which are, in turn, cut through by the alaskite aplitoid granites that belong to a still later intrusive phase.

The quartz syenite-diorites, accompanied by quartz diorites and granodiorites, are exposed in the eastern part of the Karkaraliny Mountains. They penetrate and, at the contacts, metamorphose the Devonian sedimentary and extrusive rocks and are themselves intersected by dikes of andesite-basalt porphyrites, rhyolite-trachyte porphyries and other rocks genetically associated with the Upper Paleozoic extrusives. In the adjacent areas, particularly in the Al'dzhan Mountains east of the Karkaralinsk area, the diorites penetrate the faunally characterized lower strata of the Visean stage, belonging to the Lower Carboniferous. This intrusive of quartz syenite-diorites is apparently Early Hercynian in age and corresponds to the Lower Carboniferous intrusives of the Northern Balkhash region.

The next intrusion in chronological succession is represented by the biotite granites, which are located mainly in the vicinity of Lake Pashino and extend northward into the valley of the Kendara River. At the present time erosion has exposed the upper apical parts of the intrusive, which is apparently considerably greater, on the whole, than its outcrops at the surface. Close to the biotite granites in age are the dike-like bodies of hybrid quartz syenite porphyries, granosyenite porphyries and other rocks that form a gradual transition between the granite porphyries and the porphyritic biotite granites and are, evidently, the hybridized apophyses of the biotite granite intrusive. The biotite granit[e] and dikes of hybrid rocks penetrate and metamorphose the Devonian extrusive-sedimentary rocks, the Early Hercynian quartz syenite-diorites and the Upper Paleozoic extrusives; thus it can be said that these intrusives are younger than the Early Hercynian intrusive phase.

As already stated earlier in this article, the latest formations of the intrusive complex are the alaskite and aplitoid granites. The alaskite granites penetrate all the sedimentary and extrusive Paleozoic rocks of the area, as well as the intrusives of quartz syenite-diorites and biotite granites. An excellent intrusive contact between the biotite and the alaskite granites is to be seen on the northern shore of Lake Pashin[o]; the contact, which is here exposed at the surface, extends for several kilometers and is to b[e] observed in a number of outcrops. The alaskite granites penetrate not only the biotite granites, but also the dikes of hybrid rocks; for example, north of the elevation point 1053. 2 a thick body of hybrid quartz syenite porphyries and granosyenite porphyries, elongated approximately north-south, is intersected by the alaskite granites, which break it up into a number of individual blocks.

The alaskite granites are accompanied by aplitoid granites, which are later. In the alaskite granites and the country rocks the aplitoid granites form steeply and gently dipping dikes of various thicknesses, from several cent[i]meters to several tens of meters, which strike in various directions.

According to V. K. Monich [5], the intrusive of leucocratic biotite granites and alaskites in Central Kazakhstan have absolute ages, as determined by the argon method, of about 300 and 260 million years; the age of the Karkaraly intrusives in particular is 260 million years, so that they are Late Hercynian.

2. The Intrusive of Quartz Syenite-Diorites

The main intrusion. The quartz syenite-diorite intrusives occur between the two anticlinal uplifts and are apparently associated with the syncline. As expressed in the topography, they occupy the lower parts of the relief.

The intrusives are intersected by numerous joints whose disposition shows some regularities. The most distinct are the joints with almost meridional strikes; these contain the thickest dikes. Along the steep submeridional joints there are gently dipping joints striking in the same direction, and also steeply dipping subequatorial joints parallel to the trend of the folding. Joints with northwestward and northeastward strikes are also encountered.

The intrusive is composed of medium- and fine-grained massive rocks; under the microscope these are observed to have a hypidiomorphic crystal structure and areas of monzonitic rock. The rocks consist of zonal plagioclase whose composition varies mainly from basic oligoclase No. 29 to andesine No. 50, of potassium feldspar, of ordinary hornblende with a birefringence of 0.016 to 0.017, of biotite and of light-gray quartz. The accessories are ore minerals, apatite, sphene and zircon. The secondary minerals are chlorite, secondary amphibole, epidote, saussurite, sericite, leucoxene and others.

It will be seen from the data in Table 1 that the intrusive is represented by a wide range of rocks, from quartz diorites to quartz syenite-diorites to granodiorites, with a predominance of quartz syenite-diorites, so that the intrusive as a whole has an intermediate composition. The variety of the rocks is apparently due to the assimilation by the magma of the country rocks and of the formation of the intrusives under hypabyssal conditions, thus facilitating their relatively rapid crystallization; the magma's composition was thus apparently not sufficiently evened out.

Table 1

Quantitative Proportions of Rock-forming Minerals

Thin section No.	Plagioclase	Potash feldspar	Quartz	Dark minerals	Accessories
93	50.1	11.4	12.6	21.4	4.5
108	41.0	20.0	11.2	25.5	2.5
187	62.1	16.5	11.3	7.2	2.9
253	50.4	23.0	9.8	14.2	0.5
255	48.6	24.4	12 6	13.9	0.5
310	44.6	21.9	16.3	16.3	0.9
313	46.4	15.5	15.0	20.6	2.5
327	35.1	32.1	17.8	12.4	2.6
Average	47.3	20.6	13.3	16.4	2.1

The rocks are somewhat oversaturated with silica and are rich in alkalies (Table 2). In their chemical composition they are intermediate between quartz diorites and syenites; this is in full agreement with their intermediate quantitative mineral content. The relative amounts of iron and magnesium in the light-colored components and the accessories are very similar, and the ratio of f' to m' is 1.1.

The endocontacts and dikes. The oscillations in the composition of the intrusives show a relationship to the composition of the surrounding Devonian deposits: the quartz syenite-diorites and granodiorites predominate in areas of clay shales interlayered with sandstones; at the contacts with the diabase porphyrites and their tuffs the rocks grade into quartz gabbro-diorites and quartz gabbros.

The typical quartz gabbros have a medium-grained gabbroic texture. Their composition includes: zonal plagioclase ranging from andesine No. 34 to labradorite No. 54, relicts of monoclinic pyroxene, replaced by greenish-brown hornblende associated with secondary amphibole, chloritized biotite, and quartz with even or slightly wavy extinction. The accessories are: opaque ore minerals, numerous microscopic prisms of apatite, and sphene. The secondary minerals are: amphibole, chlorite, epidote, zoisite, leucoxene and sericite.

The rocks are classes as quartz gabbro (Table 3 and 4) moderately rich in alkalies. There is typically a high titanium content and a predominance of magnesium over iron in the dark components; the ratio of f' to m' is 0.8.

The quartz syenite-diorites are cut by dikes of various compositions. The dikes, which are associated properly with the intrusives of quartz syenite-diorites, are represented by syenite-diorite porphyrites and by quartz syenite-diorite porphyrites; they occur both in the intrusives and within the Devonian rocks. Their trends are meridional, northeastward and northwestward; they are of small thicknesses, amounting to no more than a few meters.

In addition to the dikes of syenite-diorite porphyrites, the quartz syenite-diorites are intersected by dikes genetically associated with the Upper Paleozoic extrusives, and by numerous dikes of hybrid rocks related to a later intrusive phase.

3. The Intrusion of Biotite Granites

The main intrusion. The distribution of the biotite granites has been discussed above. They occur primarily in the area of Lake Pashino, probably because the biotite granites, like the intrusion of the quartz syenite-diorites, are associated with a synclinal fold that is at the present time masked by the various different plutonic rocks that have been intruded into it.

The biotite granites are characterized by excellent jointing that separated them into thin layers, reflected in the development of numerous horizontal and gently dipping fractures. Some times these granites show a coarse sort of slaty cleavage.

In addition to the gently dipping fractures, the biotite granites contain submeridional and subequatorial joint fractures that are respectively

Table 2

Chemical Compositions of Quartz Syenite-Diorites

Oxides	Specimen 108		Specimen 187	
	Wt. %	Mole quantity	Wt. %	Mole quantity
SiO_2	62.22	1036	63.49	1058
TiO_2	0.66	8	0.60	8
Al_2O_3	16.07	158	15.92	156
Fe_2O_3	2.38	15	2.29	14
FeO	2.90	40	2.33	32
MnO	0.10	1	0.08	1
MgO	2.72	68	2.24	56
CaO	4.66	33	4.17	74
Na_2O	3.85	62	3.98	65
K_2O	3.11	33	3.35	35
B_2O_3	None	None	0.02	—
F	0.03	2	None	None
H_2O^+	0.86	48	0.95	53
H_2O^-	0.08	—	0.11	—
Total	99.64	—	99.53	—

Specimen 108		Specimen 187	
$a = 13.1$	$n = 65.3$	$a = 13.8$	$n = 65.0$
$b = 10.9$	$f' = 44.6$	$b = 9.3$	$f' = 45.2$
$c = 4.3$	$m' = 42.8$	$c = 3.9$	$m' = 41.5$
$S = 77.7$	$c' = 12.6$	$S = 73.0$	$c' = 13.3$
$Q = +12.5$	$\varphi = 18.87$	$Q = +14.5$	$\varphi = 20.7$
$\frac{a}{c} = 3.0$	$t = 0.77$	$\frac{a}{c} = 3.5$	$t = 0.76$

Table 3

Per-cent Mineral Composition of Quartz Gabbro

Thin section No.	Plagioclase	Amphibole, pyroxene, chlorite (after biotite)	Quartz	Accessories
266	53.3	36.8	7.0	2.9
266 a	47.5	42.0	5.6	4.0

Table 4

Chemical Composition of Quartz Gabbro (Specimen 266)

Oxides	Wt. %	Mole quantity	
SiO_2	48.38	806	$a = 5.8$
TiO_2	1.22	15	$b = 33.4$
Al_2O_3	14.91	146	$c = 4.3$
Fe_2O_3	5.66	35	$S = 56.5$
FeO	7.01	97	$Q = -2.6$
MnO	0.22	3	$\frac{a}{c} = 1.4$
MgO	8.34	208	$n = 73.8$
CaO	9.48	170	$f' = 35.0$
Na_2O	1.93	31	$m' = 42.8$
K_2O	1.09	11	$c' = 22.2$
B_2O_3	None	None	$\varphi = 14.4$
H_2O^+	1.28	71	$t = 1.8$
H_2O^-	0.28	—	
Total	99.80	—	

perpendicular and parallel to trend of the folding, as well as diagonal fractures striking northeastward and northwestward.

The texture of the granites is fine-grained and very rarely medium-grained porphyritic. The phenocrysts consist of microperthitic microcline and of acidic plagioclase. The ground mass is made up of microperthitic microcline with indistinct microcline grating, of zonal albite-oligoclase Nos. 7-30, of small grains of quartz and of biotite. The accessories are ore minerals, apatite, sphene, rutile, zircon and monazite. The secondary minerals are chlorite, sericite, muscovite and others.

The potash feldspar-microperthite predominates sharply over plagioclase in amount (Table 5). Characteristically there are a few small growths of albite in the microperthite: the ratio of these to the potash feldspar proper within the microperthite is approximately 1 to 4.5 or 5.

The rock is supersaturated in aluminum oxide and silica and is also rich in alkali, particularly potassium oxide (Table 6). In the biotite and accessory minerals, iron predominates over magnesium, with a ratio of $\underline{f'}$ to $\underline{m'}$ of 2.7; trivalent iron and aluminum are abundant.

The endocontacts and dikes. The endocontact facies are most typically reflected in the apophyses of the intrusive that form dike-shaped bodies and dikes, varying in thickness from several hundreds of meters to 1 meter or less. On being injected into the quartz syenite-diorites diabase porphyrites and other rocks relatively low in silica, these apophyses of the granitic magma were hybridized by the assimilation of the country rocks and were crystallized with the formation of quartz syenites and quartz syenite porphyries, granosyenites and granosyenite porphyries. The thin dikes are for the most part composed of granite porphyries, since the process of assimilation in them was probably less intensive because of the more rapid cooling. There are also dike-shaped bodies of complex composition, formed of both the hybrid granosyenite porphyries and quartz syenite porphyries and of the porphyritic biotite granites, granite porphyries and quartz porphyries.

The chemical composition of the hybrid rocks

Table 5

Quantitative Mineral Composition Biotite Granites

Thin section No.	Quartz	Microperthite: Potash feldspar	Microperthite: Albite	Acidic plagioclase	Biotite	Accessories
103	33.7	53.3		10.2	1.9	0.9
264	46.4	26.3	5.4	16.9	4.0	1.0
264	38.9	38.0	8.4	10.0	3.1	1.6
501	39.5	30.7	6.4	20.8	1.9	0.7

Table 6

Chemical Composition of Porphyroidal Biotite Granite (Specimen 103)

Oxides	Wt. %	Mole quantity	
SiO_2	71.80	1196	a = 5.6
TiO_2	0.22	3	c = 1.5
Al_2O_3	14.88	146	b = 2.8
Fe_2O_3	1.03	6	S = 80.1
FeO	0.71	10	Q = +27.5
MnO	0.03	—	
MgO	0.30	8	$\frac{a}{c}$ = 10.4
CaO	1.30	23	n = 51.3
Na_2O	3.73	60	f' = 52.4
K_2O	5.41	57	m' = 19.1
B_2O_3	None	None	a' = 28.5
S	0.022	0.7	φ = 28.6
F	0.02	1.1	t = 0.25
H_2O^+	0.40	22	
H_2O^-	0.27	—	
Total	100.122	—	

that form the dikes, which was given in an earlier paper by the present writer, is briefly characterized by the following data (in weight %): SiO_2 — 65.10; TiO_2 — 0.71; Al_2O_3 — 16.82; Fe_2O_3 — 3.25; FeO — 1.14; MnO — 0.12; MgO — 0.61; CaO — 1.90; Na_2O — 4.81; K_2O — 5.22; H_2O^+ — 0.50; H_2O^- — 0.32. These rocks do not differ from the granites in their potassium content, but they have much less SiO_2 and much greater quantities of Al_2O_3, iron oxides, alkaline earths and sodium; the reason for these differences is apparently the hybridization.

The dikes and dike-shaped bodies of hybrid rocks strike in various directions, and are located in zones of faults and fractures in the surrounding rocks. Like the biotite granites, they were probably formed under hypabyssal conditions and, in some cases at least, were probably associated with Late Paleozoic extrusive activity, as suggested by V. I. Yagovkin (1955) in the case of other intrusives of similar age and composition in the Karkaralinsk area.

4. The Intrusion of the Alaskite and Aplitoid Granites

The alaskite granites. The principal mass of the alaskite granites, which forms the northwestern part of the Karkaralinsk intrusive complex, is apparently associated with the area of the axial plunge of a subequatorially trending anticline; the structure of this area was considered above, in the discussion of the tectonics of the district. The southwestern part of the massif, however, is located in the syncline, whose axial part, trending almost parallel to the equator, was in all probability composed of Upper Devonian and Upper Paleozoic deposits which by now have been almost completely replaced by the Karkaralinsk intrusive complex.

Tongues of alaskite granites that branch out eastward from the principal massif made use of the subordinate tectonic structures and are sometimes intruded into the secondary anticlinal folds. One example of such a tongue is the mass of alaskite and aplitoid granites between the Komsomolets and Kyzylshoky kolkhozes, located at the plunging axis of a subequatorial anticlinal fold.

The alaskite granites, which cut across the biotite granites in the vicinity of Lake Pashino, are frequently located along the surfaces of gently dipping joints in the biotite granites, which serve as their lower exocontact; the roots of the intrusive plunge steeply into the depths and their contacts follow the steep joints in these same biotite granites. Thus the alaskite granites in this area form irregular laccolith-shaped bodies with exposed lower contacts.

In the center of the masses of alaskite granites there is a predominance of systems of steeply dipping joints whose strikes are NE 50° and NW 320°, in addition to steep submeridional and subequatorial fractures, the latter being less distinct and often masked by the joints striking in the former directions. The gently dipping joints usually strike parallel to the contacts and dip toward the contact at angles of from 5° to 20°.

The alaskite granites have a medium-grained and less often coarse-grained aplitoid texture in which the quartz grains characteristically have equidimensional shapes. The chief minerals are quartz, more-or-less equal-sized microperthitic microcline, platy crystals of albite and oligoclase Nos. 0-24, and biotite. The microcline frequently has typical microcline cross twinning. The accessories are fine-grained ore minerals,

zircon, sphene and monazite (?). The alteration products are chlorite, sericite, muscovite, albite, abundant fluorite and others.

According to their mineral composition the granites are of the leococratic variety, rich in quartz (Table 7). The ratios of the albite to the potash feldspar proper in the microperthite are approximately 1:2 or 1:3.

Table 7

Quantitative Mineral Composition of Alaskite Granites

Thin section No.	Quartz	Microperthite		Acidic plagioclase	Biotite	Accessories
		Potash feldspar	Albite			
4a	52.2	29.0		13.2	3.6	
43	47.2	37.0		15.8	—	
1036	54.8	29.5		15.0	0.7	
1036	54.5	22.4	13.5	8.8	0.4	0.4
1036	32.8	34.7	10.9	12.3	8.0	1.3
171	36.9	34.7	18.5	8.7	0.7	0.5
272	40.5	33.2	10.7	13.3	1.7	0.7

The rocks are oversaturated with silica and rich in alkalies (Table 8). In their chemical, as well as their mineral, composition they correspond to alaskites. In the biotite and accessories there is several times more iron than magnesium, but the ratio of f' to m' is not the same (3 and 7. 5). The amount of sodium oxide by weight is less than the amount of potassium oxide in the ratio of 1 to 1. 3-1. 5; the mole ratio of sodium to potassium oxide is 1 to 0. 9 and sometimes 1 to 1.

The endocontacts of the alaskite granites. In the endocontact areas of the alaskite granite intrusives there is usually an increase in the number of fine-grained aplitoid granite dikes along with an appearance of pegmatoid segregations that form irregular lenses from several centimeters to several tens of centimeters or more in size. Sometimes the pegmatoid formations combine with the aplitoid granites to form veins. In the contact areas the alaskite granites themselves differ little from the granites in the central parts of the intrusives, and in places one may note a slight decrease in the size of the components forming them, so that the usually

Table 8

Chemical Composition of Alaskite Granites

Oxides	Specimen 4		Specimen 103b		Specimen 148		Specimen 171	
	Wt. %	Mole quantity	Wt. %	Mole quantity	Wt. %	Mole quantity	Wt. %	Mole quantity
SiO_2	78.02	1299	76.09	1263	74.38	12.38	77.06	1284
TiO_2	0.15	2	0.11	1	—	—	—	—
Al_2O_3	11.81	116	13.08	128	13.52	132	11.90	117
Fe_2O_3	0.56	4	0.18	1	1.57	—	0.98	—
FeO	0.50	7	0.28	4				
MnO	0.02	—	0.02	—	—	—	—	—
MgO	0.08	2	0.08	2	0.22	5	0.25	6
CaO	0.42	7	0.31	5	1.19	21	1.19	21
Na_2O	3.41	55	4.05	65	3.76	61	3.55	57
K_2O	4.63	49	5.42	57	5.07	54	5.22	55
B_2O_3	None	None	None	None	—	—	—	—
F	0.03	2	None	None	None	None	0.05	3
H_2O^+	0.23	13	0.23	13	0.13	7	0.26	14
H_2O^-	0.28	—	—	—	—	—	—	—
Total	100.14	—	99.89	—	99.76	—	100.46	—

Specimen 4a		Specimen 1036	
$a = 13.5$	$n = 52.9$	$a = 16$	$n = 53.3$
$c = 0.4$	$f' = 55.6$	$c = 0.3$	$f' = 60.0$
$b = 1.8$	$m' = 7.6$	$b = 0.7$	$m' = 20.0$
$S = 84.3$	$a' = 37.0$	$S = 83.0$	$c' = 20.0$
$Q = +41.2$	$\varphi = 29.6$	$Q = +33.7$	$\varphi = 20.0$
$\frac{a}{c} = 34$	$t = 0.15$	$\frac{a}{c} = 53.4$	$t = 0.07$

medium- and coarse-grained granites grade into more fine-grained varieties; more rarely the granites grade into granite porphyries of subalkaline type. But sometime there are extensive endocontact zones characterized by the fact that the alaskite granites combine with the older biotite granites to form very complex contact relationships that are described in detail below.

The aplitoid granites form dikes that are frequently located within joints in the alaskite granites, both steeply dipping and gently dipping joints. Under the microscope these have an inequidimensional-grained aplitoid texture; the grain sizes of the chief rock-forming minerals vary from 0.1 to 1 mm, but there are almost always smaller grains present, and more rarely also larger ones. Sometimes the structure is micropegmatitic.

The most important rock forming minerals are represented by microperthitic microcline, quartz, acidic plagioclase (primarily albite) and biotite. The accessories are ore minerals, zircon, sphene, and apatite. Unaltered aplitoid granites are almost never encountered, since all of them to one degree or another bear traces of post-magmatic alteration, resulting in the formation of secondary albite, muscovite fluorite and other minerals.

E. SOME LAWS GOVERNING THE FORMATION OF THE KARKARALINSK INTRUSIVE COMPLEX

1. Successive Changes in the Chemical Composition of Intrusives of Different Phases

Despite the complicated multi-phase character of the Karkaralinsk intrusive complex, its formation followed the normal succession of events, as established for a whole series of other intrusives [1, 8] and reflected in the successive replacement of the more basic by more acidic intrusives. The changes in the chemical composition of the complex is shown on the variation diagrams (Figures 2 and 3), constructed for the chief rock-forming oxides and for B_2O_3 from the chemical analyses cited earlier in this paper, and for the elements present as admixtures from numerous semiquantitative spectrum analyses.

The Early Hercynian quartz syenite-diorites and endocontact quartz gabbro, in comparison to the intrusives of later phases, contain larger amounts of the alkali earths, iron titanium, aluminum and constitutional water. The highest content of sodium is found in the quartz syenite-diorites, which also contain boron. The biotite granites are rich in potassium: they contain more than 5% K_2O. The change in the chemical composition of the later alaskite granites, in contrast to the preceding biotite granites, is reflected in a lower content of aluminum and a larger amount of silica, without any increase in the quantities of sodium or potassium; the content of the latter two elements is even lower in the biotite granites.

The successive change in the chemical composition of the intrusive complex is accompanied by a regular change in the elements present as admixtures. The Early Hercynian intrusives of more basic composition are relatively richer in admixtures of cobalt, nickel, chromium, vanadium, copper, zinc and strontium, which probably accompanied the iron, aluminum and the alkali earths in the crystallization of the magma. The potassium-rich biotite granites contain relatively large admixtures of barium. The alaskite granites show an admixture of tin, and are richer than the preceding biotite granites in beryllium and molybdenum. Characteristically, a relative predominance of certain associations of admixture elements occurs in those intrusive rocks with which the deposits of these elements are genetically associated. As examples one may cite the admixtures of nickel, chromium and copper, concentrated in the more basic rocks which, it is well known, play an active role in the formation of the industrial deposits of these elements. The acidic intrusives show higher contents of molybdenum and tin, deposits of which are usually associated with acidic igneous rocks.

2. Some Aspects of the Formation of Intrusives of Various Compositions

Quartz syenite-diorite intrusives. The formation of the quartz syenite-diorites took place among the sedimentary and extrusive rocks of the Devonian, and was apparently accompanied by intensive assimilation of these surrounding rocks. The proof of this contention is the dependence of the composition of the endocontact facies of these intrusives upon the composition of the exocontact rocks. Because of this assimilation, the Early Hercynian intrusive acquired a hybrid nature, reflected in the range of various kinds of rocks from granodiorites to quartz gabbros.

The interaction of the magma with the surrounding rocks was typically characterized by the complete or almost complete dissolution of the latter and the formation of a hybrid, but homogeneous magmatic solution with a molecular or ionic dispersion.

The assimilation of the host rocks was, naturally, accompanied by replacement of the magma that occupied their space. It is difficult to reconstruct the original composition of the magma, since the crystallization of the quartz syenite-diorites involved not only the substance of the original magma, but also the products of

assimilation. Taking account of the fact that the quartz syenite-diorites appeared in the contacts with the Devonian series of clay shales and sandstones, rich in potassium and silica, it may be supposed that the magma that took part in the assimilation and replacement of the host rocks was not of ultrabasic or alkaline composition, but contained alkali earths and iron in quantities sufficient for an intermediate magma. The formation of quartz gabbros in the contacts with the diabase porphyrites also supports this supposition.

The relationships of the dike-forming intrusives of hybrid rocks to the quartz syenite-diorites. The numerous dike-shaped bodies of granosyenites, granosyenite porphyries, quartz syenites and quartz syenite porphyries are examples of typical hybrid formations whose hybridization was due to the assimilation of more basic rocks, particularly quartz syenite-diorites. Under the microscope one may observe the contamination of the hybrid rocks by foreign material, represented by xenocrystals of zonal plagioclase, biotitized aggregates of amphibole, skeletal grains of accessory sphene and other minerals. All these show clear traces of dissolution or recrystallization and are apparently inherited from the assimilated quartz syenite-diorites.

The xenocrystals of zonal plagioclase, moreover, show growths of microperthite at the borders and form phenocrysts in the hybrid rocks (Figure 4). Sometimes, between the relict crystals of zonal plagioclase and microperthite there is an intermediate zone of antiperthite; this is probably the result of a reaction between the plagioclase xenocrystals and the granitic magma. All these phenomena are causes of the highly variable mineral composition of the rocks under discussion (Table 9).

Specimens 275 and 275a differ little from the granites, except for the character of their plagioclase (represented by zonal andesine).

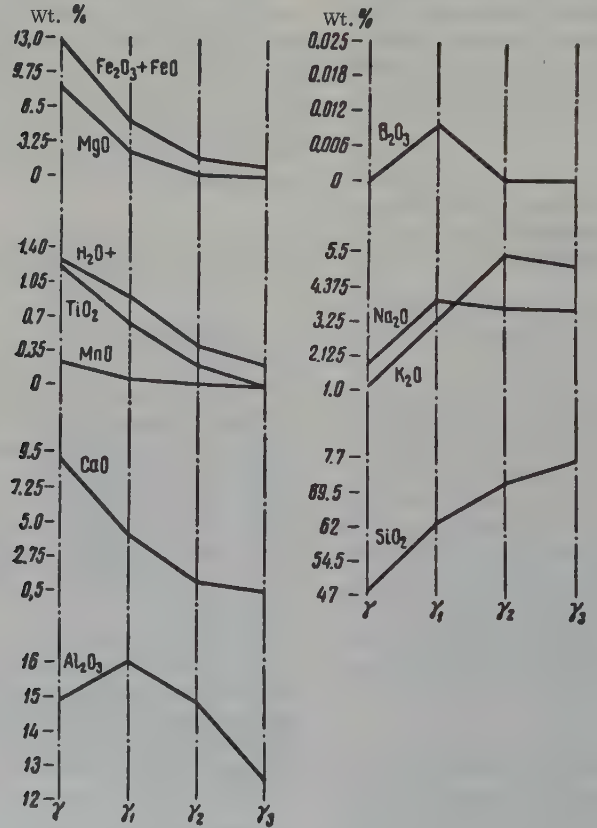

FIGURE 2. Variation diagram showing the change in the content of rock-forming oxides in the intrusive rocks.

γ - quartz gabbro, γ_1 - quartz syenite-diorites, γ_2 - biotite granites, γ_3 - alaskite granites.

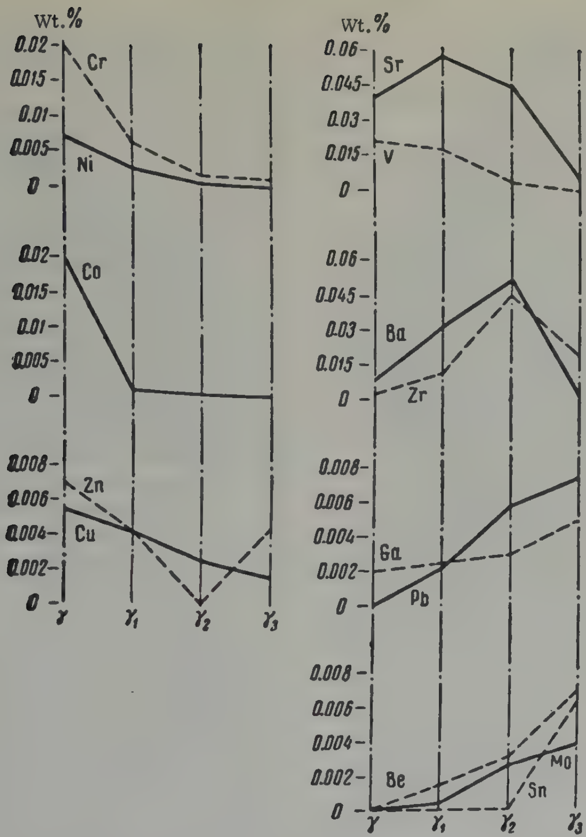

FIGURE 3. Variation diagram showing the changes in the content of minor element admixtures in the intrusive rocks.

Designations same as in Figure 2.

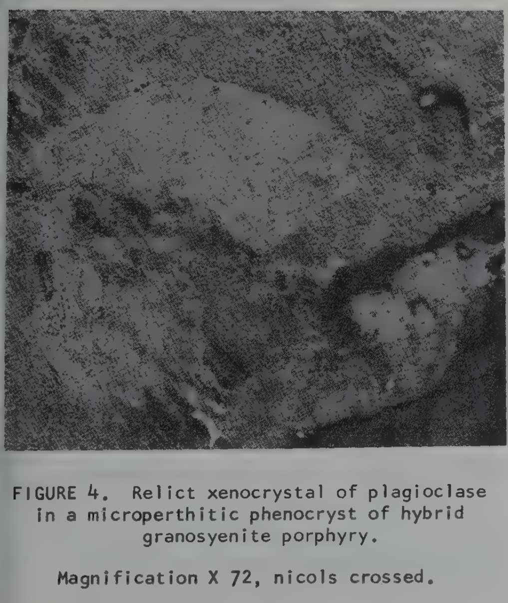

FIGURE 4. Relict xenocrystal of plagioclase in a microperthitic phenocryst of hybrid granosyenite porphyry.

Magnification X 72, nicols crossed.

The rocks of specimens 319 and 319a are very close in their mineral composition to the assimilated quartz syenite-diorites. Other rocks have hybrid compositions ranging from granosyenite to quartz syenite.

The interaction between the intrusives under consideration and the quartz syenite-diorites did not involve complete solution of the latter, since the hybrid rocks contain a number of randomly oriented microxenoliths of quartz syenite-diorites. Most likely the magma, in the contact area, was a suspension consisting of a liquid solution with many solid relict particles of plagioclase, amphibole and accessories, derived from the replaced quartz syenite-diorites, suspended within it. Comparison of the data from chemical analyses of the hybrid rocks and the quartz syenite-diorites shows that the assimilation of the latter caused the introduction mainly of potassium (since the hybrid rocks contain 5.22% K_2O as compared to 3.11-3.35% in the quartz syenite-diorites) and the removal of calcium and magnesium. In this

Table 9

Mineral Composition of Hybrid Rocks

Thin section No.	Quartz	Micro-perthite and antiperthite	Plagio-clase	Amphibole and biotite	Acces-sories
104	5.8	80.9	—	1.6	0.7
272	19.3	74.9	—	5.3	0.5
272	12.3	81.2	—	5.3	0.7
272a	14.6	80.0	—	2.6	2.8
275	28.6	61.5	5.2	2.3	2.4
275a	27.4	65.1	4.7	1.4	1.4
319	12.4	42.8	29.3	11.2	4.8
319a	13.3	41.9	27.5	12.4	4.9

process of assimilation and replacement the active role was probably played by the potassium cations, which because of their great chemical reactivity drew out the less active cations of alkali earths from the crystal lattices of the minerals composing the quartz syenite-diorites. Traces of this extraction are preserved in the hybrid rocks in the form of reaction rims of antiperthite around the plagioclase microlenoliths. Because the reaction between the potassium-rich magma and the quartz syenite-diorite host rocks did not go to completion, the hybrid rocks after their crystallization retain clear signs of contamination.

The relationships between the alaskite granite intrusives and the biotite granites. The alaskite granites, which cut through the earlier biotite granites, sometimes form sharp and even contacts with the latter (Figure 5) or else cut across them in the form of thin tongues, veinlets (Figure 6) or other larger injections. But in a number of cases there are gradational contacts between the alaskite and biotite granites, creating a false impression that the two granites are of the same age. The contact zone here moreover contains granites of mixed composition, within which the microscope reveals relict minerals from the biotite granites and newly formed quartz and feldspar characteristic of the alaskite granites (Figure 7). In the case of these associations of minerals of different ages, the biotite, feldspar and quartz from the biotite granites are typically dissolved and replaced by the later grains of quartz, microcline and acidic plagioclase from the alaskite granites, which contain relict traces of the replaced minerals in the form of inclusions.

Under the microscope these two mineral associations of different ages are seen to differ sharply in the grain sizes of their components: the relict aggregates of quartz and feldspar from the biotite granites consist of individual crystals usually about 1 mm in size; the later quartz and feldspar grains genetically associated with the alaskite granites considerably exceed this size, reaching the size of 2 to 3 mm or more. Moreover the relict minerals differ from the newly formed ones in their habit; this is particularly clearly seen in the potassium feldspars: the potash feldspar crystals from the alaskite granites are characterized by more complete microcline network, with abundant and large microperthitic growths of albite, whose quantity, as already mentioned above, relative to the amount of potash feldspar proper is in the ratio of 1 to 2 or 1 to 3, whereas in the microperthite of the biotite granites this ratio is 1 to 4.5 or 1 to 5, and the albite growths are much thinner; the optical axes in both the potash feldspars are variable, but in the potassium feldspar genetically connected with the alaskite granites they range from -68 to -84° (specimens 503, 503a), whereas in the potash feldspar from the biotite granites the optical axes vary from -48° to -68° (specimens 103, 503). This means that the less oriented potash feldspar of the biotite granite was replaced by the more highly oriented potash feldspar of the alaskite granites.

These manifestations of the contact interaction between the alaskite and biotite granites, with the formation of a zone of mixed granites, were produced during the magmatic stage of intrusion of the alaskite granites, inasmuch as the newly formed minerals within the biotite granites completely resemble the minerals crystallized during the magmatic period in the intrusion of the alaskite granites. In addition, there was apparently an infiltration of the magma along disjunctive dislocations, accompanied by incomplete recrystallization of the biotite granites and

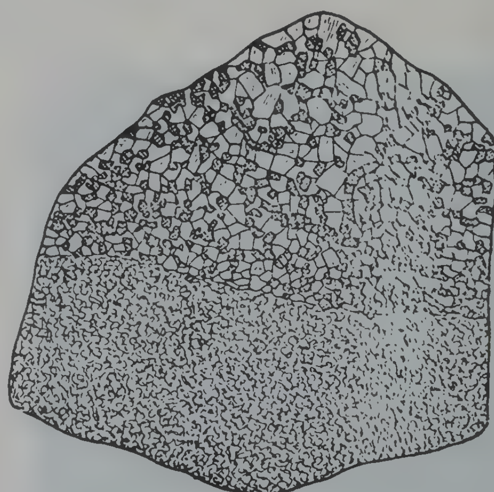

FIGURE 5. Sharp contact between coarse-grained alaskite granite and fine-grained biotite granite.

Sketch of polished surface of specimen, 3/5 natural size.

FIGURE 6. Veinlet of alaskite granite in fine-grained biotite granite.

Sketch of specimen, 3/5 natural size.

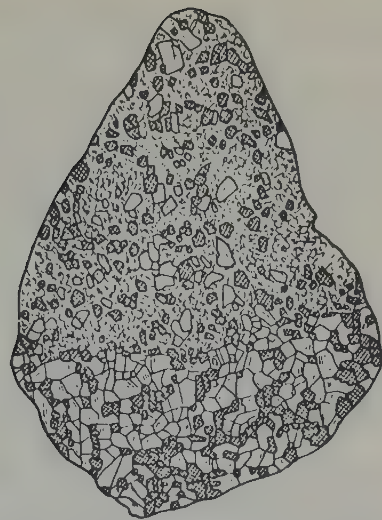

FIGURE 7. Replacement of fine-grained biotite granite by large feldspar and quartz crystals at the contact with coarse-grained alaskite granite.

Sketch of polished surface of geologic specimen, 3/5 natural size.

etasomatic replacement of the latter in the olid state by the mineral associations typo-orphic of the alaskite granites. This replace-ent should probably be regarded as a process alaskitization of the biotite granites, having eatures in common with the process of graniti-ation in certain Paleozoic intrusive complexes 2]. Apparently no alkalies were introduced uring the alaskitization of the biotite granites, owever, as may be judged by a comparison of e chemical compositions of the biotite and laskite granites. The enrichment of the gran-es in silica during their alaskitization and the ecrease in the content of the majority of the ther principal components suggests that the agmatic solution that played the active part in e replacement of the biotite granites was ex-remely rich in silicic acid, with some excess water. The infiltration of the solution into e biotite granites and its participation in the rocess of their recrystallization were probably cilitated by the loss of excess water, by the ssolution of the silica and by the precipitation quartz from the residual melt, thus termina-ng the crystallization of the purely intrusive asses. The later aplite-pegmatite and hydro-ermal processes were locally developed post-agmatic events superimposed upon the already rmed intrusive massif of alaskite granites and eir exocontacts.

CONCLUSION

The formation of the Karkaralinsk intrusive omplex in the Hercynian igneous cycle took lace on the northern margin of the Paleozoic zhungara-Balkhash geosyncline. The Early ercynian intrusives formed in the Middle Paleo-oic structural stage are represented by a gamut f rocks from granodiorites to quartz gabbros, ith a predominance of quartz syenite-diorites.

The later intrusives, located both in the Middle Paleozoic and the Upper Paleozoic structural stage, are represented by potassium-rich biotite granites and hybrid rocks formed during the closing stages of the folding. The Late Hercynian intrusives, formed after the folding and during the pre-platform stage, are represented by silica-rich alaskite granites.

Observations in the areas of the contacts of these intrusives show that their formation was accompanied by replacement of the host rocks by the magma, despite the fact that many of the intrusives were formed at small depths. The replacement processes differed, depending on the composition of the intrusion, and was manifested in the assimilation of the exocontact rocks or their metasomatic replacement in the solid state.

The composition of the magmatic solutions that played the active part in the replacement processes apparently changed successively, from alkali earth and potassium-rich in the early stages to ultra-acidic in the last stages of the formation of the intrusive complex.

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STYLOLITES[1]

by

G. I. Bushinskiy

This article describes the morphology, terminology, occurrence and origin of stylolites, as well as the methods of studying them; the significance of stylolites in lithology, tectonics, in the study of oil and gas collectors and of economic mineral deposits is noted.

* * * * *

In the entire problem of stylolites, the least thoroughly developed aspect is the methods of studying them. As a result of inadequate observations, a large number of erroneous hypotheses have been put forth regarding the origin of stylolites, and their theoretical and practical significance has also not been properly understood. This article will attempt to remedy this deficiency.

The history of the study of stylolites and of the views regarding their origin may be found in papers by M. S. Shvetsov, G. I. Teodorovich, V. N. Kholodov, K. K. Zelenov and D. V. Nalivkin [5, 11, 17, 18, 20, 22].

D. V. Nalivkin [11] defines stylolites as "peculiar columnar formations occurring in limestones, and more rarely in sandstones, quartzites and clay shales. The name "stylolite" is derived from the Greek words "stylos" — column, and "lithos" — rock, and thus indicates a column-like rock. The initial form of a stylolite is the so-called "denticular seam". Each dentation in such a seam or suture represents a stylolite in the initial stage of its growth."

In the course of his geological investigations of the Russian platform, the Caucasus, the Urals, Siberia, and Soviet Far East and China, the author of the present article has had many occasions to observe stylolites of various kinds. An extraordinary abundance of stylolites may be seen in the Crimean, Altay and Ural marbles, with which the stations of the Leningrad and Moscow subways have been faced. The marble sheets that face the interiors of buildings in Moscow, Leningrad, L'vov, Sverdlovsk and other cities also contain numerous stylolites.

Various stylolites are also encountered in studying thin sections of carbonate and silicate rocks, quartzites and phosphorites under the microscope.

Stylolitic formations are most frequently observed in the cross section of a bed. Here they have the appearance of a denticulated seam or suture, along which the layers or parts of the rock are joined. The stylolite seam is marked by a thin clay layer; if this is very thin, it is called a film, and if it is thicker, it is an interbed. But the latter name is incorrect, since the clay layer does not always coincide with the bedding, but frequently intersects the beds and arises as a result of secondary changes in the rock, and not of the successive layering of the carbonate or argillaceous sediments.

In studying stylolites, it is necessary to distinguish between: 1) the stylolite surface and 2) its elements — the columns, teeth, hummocks and the depressions between these projections. In the perpendicular section, the stylolite surface looks like a toothed seam, and if the teeth are smoothed out, it is called a median surface. In plan view, the stylolitic surface is somewhat wavy and is covered with numerous scattered projections, which are the stylolites. According to the shape of the stylolitic elements, stylolites are classified as columnar, denticular or hummocky (Figure 1).

Stylolites of the first category have the form of columns, needles, bars and truncated cones of almost columnar shape (Figure 2). The tops of the columns are fairly flat and their shapes are circular in cross section, but they may be irregular or sometimes quite strongly elongated. Such elongated columns often approach bar stylolites in shape.

Very large bar stylolites have been observed by A. V. Kopeliovich in the Silurian limestones

[1]O stilolitakh, (pp. 39-57).

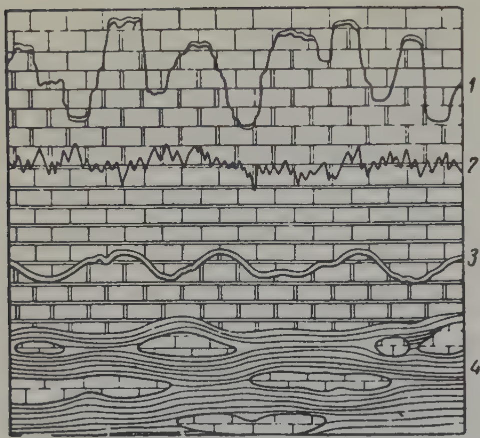

FIGURE 1. Types of stylolites.

1 - columnar; 2 - denticular; 3 - hummocky; 4 - nodular structure closely resembling stylolites. 3/5 natural size.

of Podolia. The lateral surfaces of the columns there have longitudinal grooves or striations. These columnar stylolites are accompanied by denticular stylolites, and occur primarily in pure carbonate rocks. The clay layer in these columnar stylolites is usually very thin, or altogether lacking. At the tops of the colums the clay layer forms caps or small hoods; the longer the column, the thicker the cap. A small clay layer is also observed along the furrows in the sides of the columns, and fibres of serpentine occur in the dolomites of the Kashirian stage at Oka [1].

The denticular or conical-denticular stylolites have teeth in the form of cones or shapes approaching cones. The walls of the cones have longitudinal grooves and usually bear smaller teeth (Figures 3 and 4). The angle formed at the apex of the cone varies. The stylolites in pure carbonate rocks generally have sharp teeth, whereas those in more clayey rocks have less sharp or even blunt teeth. The clay layer is almost always present and readily discernible. Denticular stylolites are very common in both pure and argillaceous carbonate rocks, and also in quartzites, haline and siliceous rocks and phosphorites.

Hummocky stylolites, or parastylolites in M. S. Shvetsov's terminology [22], have the shape of hummocks or projections some 1 to 8 cm high, and are usually accompanied by conical-denticular stylolites. The clay layer in this case reaches a thickness of 5 cm, and is consequently a true clay interbed. When the hummocks have gently sloping sides and the accompanying clay layer is relatively flat, like a normally deposited sedimentary layer of clay, it is difficult to prove that the clay layer and the hummocks are, in fact, stylolites. The most important indication of their stylolitic nature is the rapid lateral transition of the clay layer into the limestone. It is possible that many thin clay interbeds that wedge out rapidly in carbonate rocks have been formed in a manner similar to the clay layers of stylolites.

Hummocky stylolites occur in the argillaceous limestones of the Cambrian system in Siberia, in the Silurian of the Baltic region, in the Hauterivian of Kislovodsk, and in the gypsiferous rocks of the Urals and the Carpathian regions. In

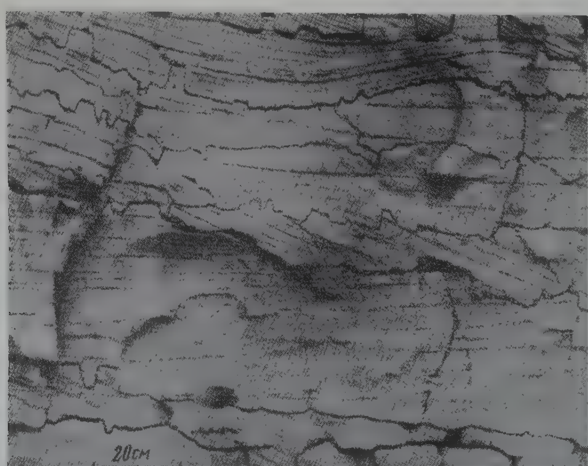

FIGURE 2. Stylolites in Upper Cambrian bedded limestones.

Outcrop along the Tyung River, a left tributary of the Vilyuy River. As a result of the distinct bedding, one can see the injection of adjacent limestone layers into each other along the stylolitic seams.

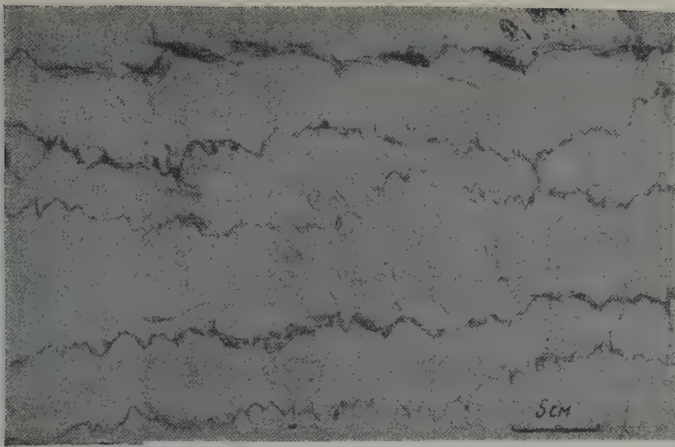

FIGURE 3. Denticular stylolitic surface in cross section.

Hummocky stylolites appear at upper left. Turonian limestone, Kislovodsk.

many places in Leningrad one may see Silurian limestones with denticular and hummocky stylolites in the foundations and wall of buildings and in the pavements.

M. S. Shvetsov's statement that nodular macrotexture is close to parastylolites is supported by the present writer's observations as well. In folded regions, as a matter of fact, argillaceous limestones frequently contain short lenses of marl or clay shale containing lenticular pieces of limestones that are circular or elongated in the direction of the bedding. These pieces are disconnected or very slightly connected along the bedding, and in places grade into layers with narrow argillaceous bridges between them, or further into a continuous mass of limestone. The lateral surfaces of the limestone pieces sometimes show grooves and smooth planes recalling slicken sides. In many cases the boundary between the pieces of limestone and the clay shale surrounding them is not sharp.

Nodular macrotexture has been observed by this writer in the intensively dislocated Silurian rocks exposed on the shore of the Sea of Okhotsk near the port of Ayan. Here the layers of black graphitic dolomitized limestones contain narrow bridges, and grade laterally into lenses and pieces of limestone enclosed within the black clay shale. Moreover the thicker parts of the lenses in one layer correspond in position to the narrow necks in the adjacent bed.

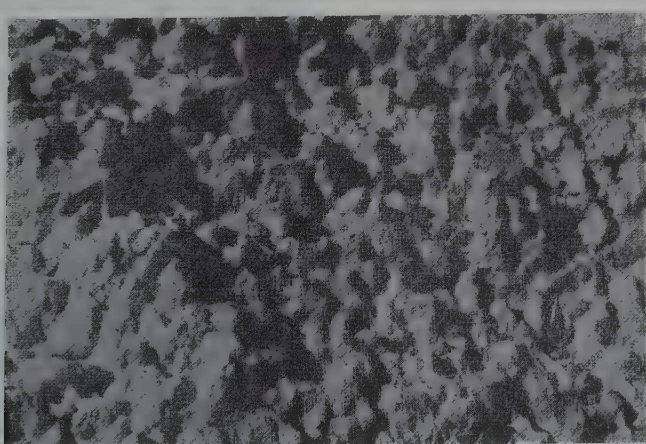

FIGURE 4. Denticular stylolitic surface as seen from above.

Turonian limestone, Kislovodsk.

M. S. Shvetsov [22] considers such formations to be the remains of continuous layers of limestone that once were interbedded with the clay rock or marl and subsequently underwent partial or complete solution. Even the masses of limestone that occur discretely along the trend of folded regions are considered by M. S. Shvetsov to be formations of similar type. Certain isolated occurrences of gypsum may also be genetically related to these formations.

The height of the stylolite columns and teeth varies greatly, from hundredths of a millimeter to as much as 90 centimeters. Sya Ban-dun [15], who has studied the stylolites in Southeastern China, divides them according to their size into four groups: large, 5-10 cm; medium, 0.5-5 cm; small, 0.1-0.5 cm and microstylolites, smaller than 0.1 cm. An almost identical classification of stylolites by size was published at about the same time by the author of the present article [3]; the latter was based upon the study of stylolites in the Soviet Union. The only difference between the Chinese and this classification of stylolites lies in the fact that the latter classes as large stylolites those with teeth higher than 20 mm.

Large stylolites have been observed only in carbonate rocks. Their denticulations are, as a rule, of columnar shape. Stylolites of medium size are also characteristic only of carbonate rocks. Small stylolites and microstylolites occur both in carbonate and in other hard rocks. Microstylolites are particularly common in ancient quartzites and sandstones. Medium and small stylolites for the most part are of denticular form, and columnar shapes are rarely encountered among them.

In regard to the median surface of the stylolite seam, the stylolites themselves are usually normal and rarely at an oblique angle to it. The sides of the columns and denticulations are more-or-less parallel to each other, even when the stylolite seam changes its inclination. In the latter case inclines stylolites are formed, the teeth of which are oblique to the median stylolite surface.

If large stylolites are developed in a thin-bedded rock, the denticulations may intersect the bedding surfaces and form what is called "ruin marble". The stylolitic teeth and columns of one such marble from Florence, slightly deformed, do actually resemble a picture of the half-ruined towers and castles around this city [11, 27]. Stylolite seams are for the most part quite even, and somewhat undulating. The median stylolite surfaces of the same generation are approximately parallel to each other.

In the Tayzskian marls of the Campanian stage in the Transcaucasus, the stylolite surfaces form hemispherical projections some 10-15 cm high. The slopes of these projections bear very small inclines stylolites parallel to the straight stylolites located along the even surfaces of the seams. Here and in other places branching stylolite seams are quite frequently encountered (Figures 5 and 6). The branches diverge only slightly from the direction of the main seams, so that the principle or parallel seams is, in essence, maintained.

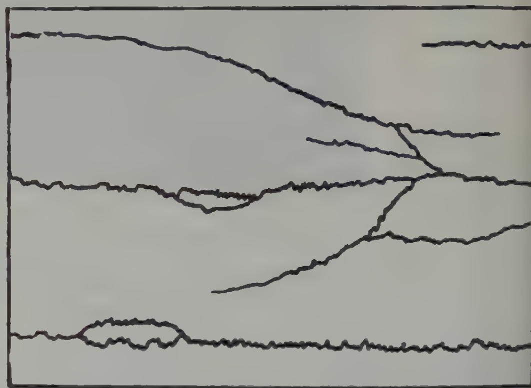

FIGURE 5. Branching stylolite seams.

Campanian marl, Tauz village, Azerbaydzhan. 3/4 natural size.

A Turonian limestone in the vicinity of Sukhumi in the Caucasus contains two systems of stylolite seams. Figure 7 shows the seams of the first system running from left to right, some of them cutting across the vertical seams of the other system and others ending at the vertical seams. The seams of the first system are roughly parallel to the bedding, and probably originated as a result of solution under the pressure of the load of overlying rocks; the second system was produced by tectonic pressure during the time of folding or afterwards by the pressure of the vertical load on the steeply dipping beds.

In determining the time of formation of one system of stylolite seams relative to the other one cannot use the same methods as in the case

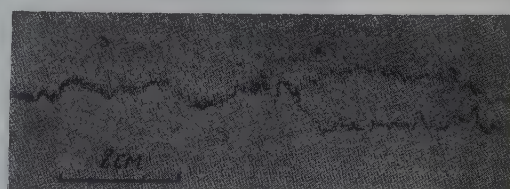

FIGURE 6. Branching stylolite seam.

Turonian limestone, Sukhumi. Natural size.

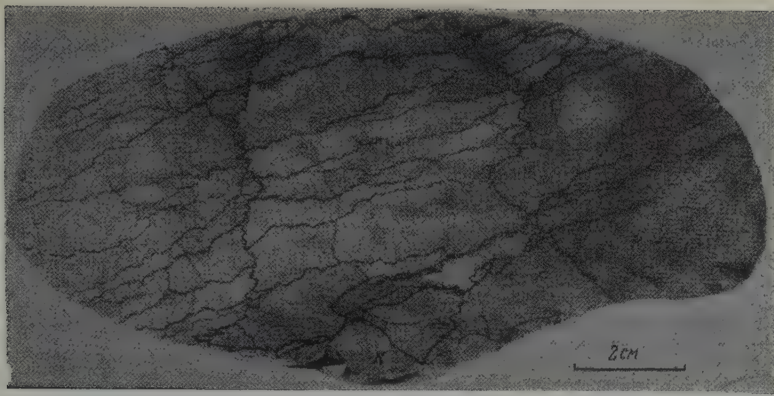

FIGURE 7. Intersecting stylolite seams.

Turonian limestone, Sukhumi. K - chert concretions. 3/4 natural size.

of mineral veins or veinlets. Actually, during the formation of successive systems of seams and the solution of parts of the limestone along such seams, there is a displacement of the intersection somewhat like the faulting along a jagged line. The magnitude of this displacement depends on the angle at which the systems intersect, on the thickness of the dissolved layer of limestone and on a possible primary tectonic displacement along a fracture that subsequently developed into a stylolite seam. Figure 8 shows the apparent displacement of an oblique calcite vein resulting from the solution of the limestone along a stylolite seam.

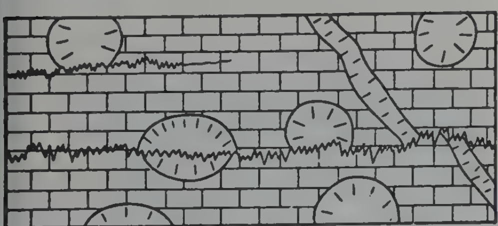

FIGURE 8. Stylolite seams intersecting coral calyzes and a calcite veinlet.

Upper Jurassic limestone, Crimea; 1/2 natural size.

Stylolite seams are either parallel to the beds or else intersect them. Intersecting stylolite seams have been described repeatedly in the literature [19, 20, 24, 31, 32]. When the beds are horizontal, only seams parallel to the bedding have been observed. In dislocated beds, there are several possible cases. When the stylolite seams are parallel to the bedding, the denticulations may be either perpendicular or oblique to the beds. In the first case, the stylolites have evidently developed as a result of the pressure caused by the weight of the overlying rocks when the bedding is horizontal; in the second case they are developed when the beds are inclined or under lateral pressure.

In studying the origin of stylolites it is very important to pay attention to the relationships between the stylolite elements and the elements of the rock structure — pebbles, sand grains, organic remains, concretions, etc. Between solid adjacent elements held together by a friable cement, the stylolite seams are limited to the area of contact between the solid elements and do not continue into the cement (Figure 9), but if the cement is also hard, the seams extend into it as well (Figures 8 and 10).

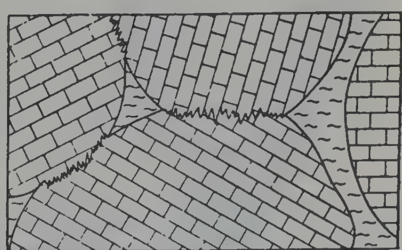

FIGURE 9. Conglomerate composed of limestone pebbles in a marl cement. Stylolite seams have developed at the contacts between the pebbles.

Devonian, Northern Urals 1/2 natural size.

Excellent stylolite seams are not infrequently

found between pebbles and pieces of limestone. The Ukrainian and Slovakian Carpathians contain Upper Jurassic conglomerates consisting of white limestone pebbles contained in a brown cement. From this conglomerate are prepared the polished slabs that may be seen in many buildings in the city of L'vov. At the contacts between adjacent pebbles in this conglomerate there are well developed fine-toothed stylolite

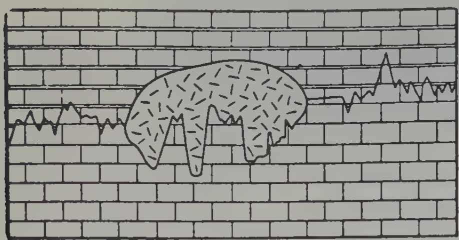

FIGURE 10. Pebble introduced into a limestone along a stylolite seam.

Sarmatian limestone, Podolia. 1/2 natural size.

FIGURE 11. "Symmetricl stylolite", a chert concretion from chalk.

Turonian, Donets Basin.

seams occur in the Devonian limestone conglomerates of the Northern Urals (see Figure 9).

Symmetrical stylolites have been described by E. Kraus and A. Kumn [28, 29]. These are stylolitic striations on the lateral surfaces of pyrite concretions occurring in a white limestone of the Lower Zechsteinian in the southern part of Latvia. The present author has seen similar striations on the surfaces of yellow chert concretions contained in the Turonian graphic chalk of the Donets Basin (Figure 11), [2]. E. Kraus has put forth two suggestions for the origin of symmetrical stylolites: 1) the striations were formed during the movement of the concretions through the ooze under the force of gravity, and 2) symmetrical stylolites are the result of the pressure caused by the crystallization of the concretions in the initial stages of diagenesis.

Since charts are amorphous, the supposition that pressure produced by crystallization is significant in the formation of stylolites falls through. The striations on the sides of the concretions may be explained by contraction movements of the surrounding mass relative to the harder concretions. This is apparently the reason for the symmetrical placement of the striations on the lower and upper parts of the lateral surfaces of the concretions.

When hollow geodes with outer crusts of calcite flow together toward the center, at the contacts between the fragments of the crust, at the points of greatest contraction, stylolite seams also appear (Figure 12).

The Middle Cambrian system in China contains widespread conglomerates consisting of flat pebbles of limestone or marl. In cross section these pebbles look like fat worms or bamboo leaves, so that the conglomerate has been given the names "wormstone" and "bamboo-leaf limestone". Figure 13 shows that the contacts between many of the pebbles take the form of denticular stylolite seams, and that the vertical pebbles form wedge-like projections into the horizontal pebbles beneath.

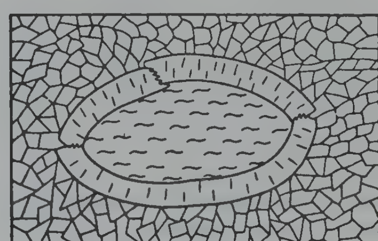

FIGURE 12. Collapsed geode surrounded by a crust of white calcite.

Stylolite seams occur where the parts of the cryst have merged into each other. Silurian limestone, Lower Tagil, Urals.

P. Herbert and R. Young [26], in mineralized dolomite breccias have observed stylolite seams cutting veinlets of sphalerite, and have come to the conclusion that these stylolites could not be primary ones.

FIGURE 13. Flat-pebble limestone conglomerate with the pebbles intersecting along stylolites.

Middle Cambrian, Mt. Syuychzhou, China.

Apart from carbonate rocks, stylolites are also found in sandstones and quartzites, in jaspers and siliceous slates, in phosphorites, bauxites, gypsum, anhydrite and haline rocks.

A. V. Kopeliovich [6-9] has distinguished the following solution structures in the sandy rocks of the Mogilev formation of Podolia: 1) microstylolites, 2) adjustments of shapes (conformations), and 3) invasions (incorporations). Among the microstylolites he distinguishes a) microstylolite surfaces of considerable extent, b) microstylolite structures at the contacts between grains, and c) microstylolites within grains. The films of matter that occur along the seams in some places represent the insoluble residue of the sandstone, and in other places the formation of new micaceous minerals.

In essence, all the microscopic solution structures observed by A. V. Kopeliovich also occur among larger stylolites in carbonate and other rocks. Microstylolites resemble macrostylolites in their forms; they also contain solumns, denticulations and hummocks.

The solution structures distinguished by A. V. Kopeliovich may be considered in a broader sense as relationships between stylolites and the structural elements of the rock — not only sandstones, but also the pebbles, organic remains, limestone fragments in breccias, veinlets, geodes and concretions. Between these elements there are also stylolitic forms of contact between the grains (see Figure 8); stylolites frequently intersect these elements, as do the microstylolites within the grains. Macrostructures reflecting adjustment of shapes occur in the Cambrian brecciated marbles of the Altay (Figure 14).

A. V. Kopeliovich's paper does not make sufficiently clear the principles by which he distinguishes various structures. Above all, he does not always rigorously and consistently use his terms "microstylolites" and "microstylolitic surface", inasmuch as the former may also refer to the individual denticulations. If a microstylolite is a surface that has the form of a series of teeth, then the shape-adaptation structure is not an independent structure, but only a particular case of a stylolitic surface and corresponds to the intergranular contact form, as defined by him. In other words, stylolite surfaces and intergranular relationships are phenomena of different orders and must not be confused with each other, as if there were a gradation between them.

A. V. Kopeliovich has shown convincingly that the solidification of sandstones and decrease in their porosity during the course of their epigenesis occurs mainly by solution of parts of the sand grains, and not by their plastic deformation. As a result of this process, the structure of an arkosic sandstone comes to resemble that of a granite. Moreover the solution of the quartz, feldspar and dark minerals that occurs during stylolitization is accompanied by their regeneration and by the formation of new minerals, primarily sericite and hydromica.

In siliceous shales and jaspers the stylolite denticulations, as in sandstones, are usually of microscopic size, rarely reaching the size of 2 to 3 mm. An example may be found in the siliceous shales of the Lesser Karatau Range, which occur at the base of the Middle Cambrian. These consist of a microaggregate of quartz and chalcedony and contain scattered microstylolite surfaces. In thin section one may see the

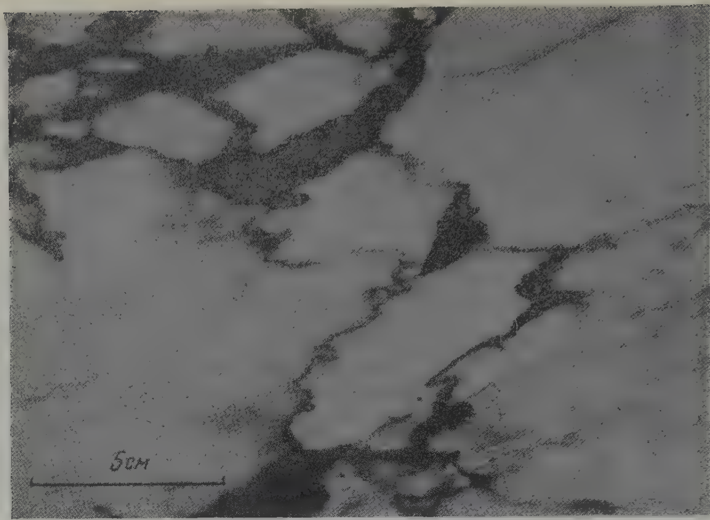

FIGURE 14. Stylolitic breccia -- a structure of shape adjustment.

Cambrian marble, Altay.

stylolitic seam and relicts of amorphous phosphate grains, which are more or less replaced by quartz (Figure 15). There has been an accumulation of phosphatic material along the stylolite seam. During the formation of the stylolites the phosphate was apparently more stable than the quartz, and silicification of the phosphate grains went on up to the beginning of this process, probably at the beginning of epigenesis.

Microstylolite surfaces occur comparatively rarely in the Karatau phosphorite. Figure 16 shows that at the contacts with the microstylolite seams the phosphate grains are partially dissolved.

The phosphorites of the Kayyan deposit, in addition to denticular stylolites, also contain flat solution surfaces. Such surfaces as the one shown in Figure 17 may be explained by attrition and scouring away of the now missing parts of the grains. But since such hemispherical grains, with their flat faces turned toward the solution surface, are also found on the other side of this surface, the explanation of the formation by attrition falls through. This phosphorite deposit also contains denticular microstylolite surfaces, which are in places so highly developed that one may suppose that no less than half of the phosphorite layer has been dissolved away.

Microstylolite surfaces are sometimes found in the Devonian bauxites of the Northern Urals.

Hummocky stylolites and parastylolites occur in gypsum and anhydrites, whereas denticular stylolites are more rarely encountered in these rocks.

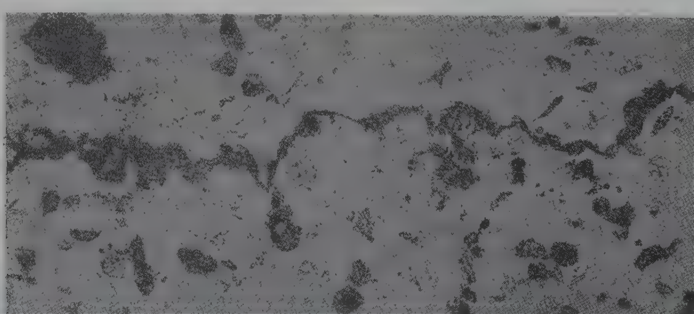

FIGURE 15. A microstylolite seam in a siliceous shale composed of amorphous calcium phosphate.

Middle Cambrian, Karatau. Magnification X 54.

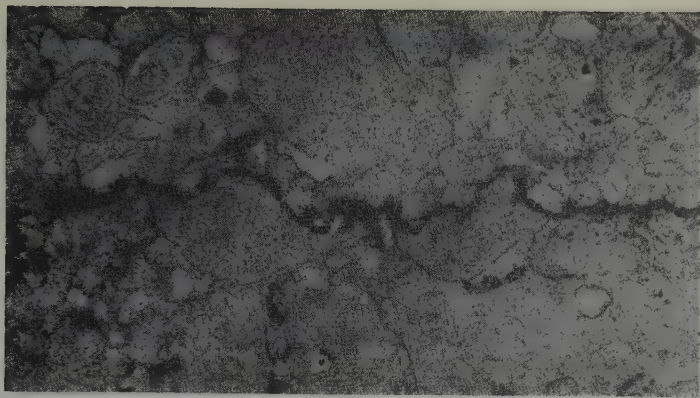

FIGURE 16. A microstylolite seam in a granular phosphorite.

The phosphorite grains along the seam are partially dissolved. Middle Cambrian, Karatau. Magnification X 54.

Conical denticular stylolites (Figure 18) have been observed in the haline deposits of the Solikamsk region. They are difficult to discern in a homogeneous salt, but are quite clearly seen in salt rocks of variegated composition.

According to the facts cited above, the morphology of stylolitic formations is characterized by three peculiarities.

1. The teeth in stylolites of the same generation are approximately parallel to each other, as well as to the direction of the pressure that caused the formation of the stylolites.

2. The parts of the rocks are readily separated along the stylolite seams; this means that stylolites do not have closed or interlocking structures. This is fully understandable, since unidirectional pressure, meaning also the formation of stylolites, cannot occur in a closed structure.

3. The median stylolite surfaces of the same generation, with straight-toothed stylolites, are roughly parallel to each other and perpendicular to the direction of the pressure.

These characteristic features of stylolites may be called the stylolite laws.

The magnitude of the pressure required to form stylolites has still not been established. A. V. Kopeliovich mentions the presence of sutures in sandstones, beginning at the depth of approximately 1000 m. According to the present writer's observations, in the Sarmatian limestones of the Dnestr River area stylolites begin

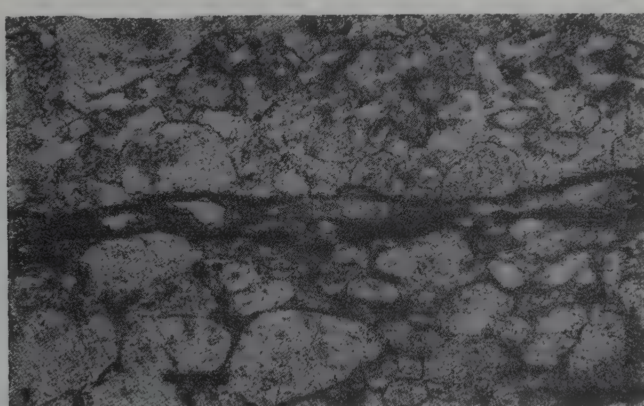

FIGURE 17. A flat solution surface in a granular phosphorite.

Half-dissolved grains may be seen at the contact of this surface. Upper Sinian, Middle phosphorite stratum of the Kayyan deposit Guychzhou province, China. Magnification X 54.

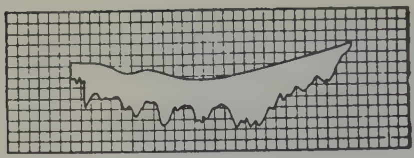

FIGURE 18. Conical denticular stylolites in haline rock: relict of red sylvite in gray halite.

1/2 natural size. Upper Permian, Solikamsk.

to appear at the depth of about 90 m, close to the level of the water table. Below this level the hydrostatic pressure increases with the depth, and the unidirectional pressure caused by the weight of the rocks should correspond to this magnitude. This circumstance, as well as ancient erosion, must be taken into account in determining the magnitude of the pressure caused by the weight of the rocks that is required for stylolites to be formed.

Extremely varied theories have been suggested in regard to the origin of stylolites: organic, crystallizational, erosional, gaseous, bituminous, pressure, pressure-solution, pressure of desiccation and karst processes. The theory of solution under pressure was suggested by T. Fuchs [24], on the basis of his study of stylolites parallel to and intersecting the bedding. His hypothesis has received many new confirmations and at the present time has the greatest number of adherents. Perhaps the most convincing observations in support of this theory are to be found in a paper by G. Wagner [32]. After studying the relationships between stylolite surfaces and organic remains, he came to the conclusion that the missing parts of the latter were dissolved out in the solid rock under pressure.

It is very difficult to find facts indicating the origin of the stylolites developed in a homogeneous rock, but in a heterogeneous rocks that contains structural elements whose sizes exceed the thickness of the stylolite seam this is much easier. The formation of the stylolites, that occur between solid pebbles, and grains, organic remains and limestone fragments (see Figures 2, 8-12) can hardly be explained except by solution of the solid rock under pressure. When the stylolite seams are developed at the contact between rocks of different kinds or else intersect grains of quartz, phosphorite, bauxite, oölites, mollusk or brachiopod tests, calcareous algae and corals, at the contact or intersection there is always a deficiency of the mass because of its solution (see Figures 9, 16, 19). From the length of the seams or by the amount of material that is lacking in organic remains and oölites whose initial dimensions are known, it is also possible to determine the amount of substance that has been dissolved.

The introduction into a limestone of shells that cap the apex of a stylolite column has been described by M. S. Shvetsov [22] and other geologists. The stylolite seams in the Turonian limestones in the vicinity of Sukhumi follow the contours of chert concretions, but do not cut through them (see Figure 7).

According to E. Riecke's principle [30], the solubility of a rock at points of increased pressure is considerably greater than its solubility in the surrounding water with hydrostatic pressure. E. Riecke has written that the crystalline grains of the rock, between which the solution saturated with its component parts circulates along the capillaries, are dissolved if the rock is subjected to unidirectional pressure, and in places where the pressure is lower the matter in solution is again deposited in the pores between the grains. This principle is confirmed by the fact that the stylolite seams developed at the contacts between the grains, pebbles, fragments and organic remains do not continue into the rock, if the latter was unconsolidated and shrank at the time of formation of the stylolites. In quartzites the silica that was dissolved in the stylolitization of the quartz grains is frequently deposited in those very places, in the form of borders regenerated of quartz. Consequently, the essence of the theory of pressure-solution may be formulated as follows: stylolites are formed as a result of solution of the solid rock at points of concentrated pressure.

One more rule follows from E. Riecke's principle and the observations that have been made: if through the fractures in a rock subjected to unidirectional pressure there is a circulation of solutions undersaturated with the components of this rock, at the points of increased pressure the material is dissolved more rapidly and is carried beyond the area of undersaturation.

This rule is confirmed by the fact that the amount of material dissolved in stylolitization often greatly exceeds the amount of material deposited in the pores of the rock.

The example of the formation of stylolites and the fillings in stylolite seams confirms the well known rule that oriented pressure is an independent factor in physico-chemical equilibrium. This confirmation is to be found in the selective solution first of the soluble salts, then gypsum, anhydrite and the carbonates and in the accumulation of various silicates, organic matter, pyrite and other minerals that form the filling of the stylolite seams. As the pressure increases, for example in the stylolitization of a sandstone, the equilibrium is displaced; not only the carbonates, but also the quartz, the skeletal silicates and the phosphates are dissolved. In this process parts of the quartz grains, which are in contact with other grains and are subjected to increased pressure at the contact

areas, undergo solution with the formation of stylolitic structures, while the dissolved silica is deposited at the same place in the pores, where there is hydrostatic pressure lower than the pressure at the contacts between the grains. Nevertheless certain authors deny the significance of stress as a physico-chemical factor in equilibrium. But our information indicates that this denial should not be extended to the entire range of temperatures and pressures.

In 1945 and 1950 G. I. Teodorovich [16, 17] proposed the hypothesis that "sutural-stylolite surfaces, or rather their embryonic forms, are usually surfaces of solution associated with karst in newly formed and consolidated calcareous deposits that have been temporarily raised up from below the water level" [17, p. 169]. Further on he also stated that in addition to subaerial, there may also have been subaqueous solution in waters with low pH, during the brief times when the water in bays of the sea or in lagoons became fresher. The development of the stylolites continued later on in the solidified deposits as a result of the circulation of waters and the dissolution of the overlying limestone.

A similar idea was suggested about 100 years ago by Plieninger and, considerably later, by V. P. Maslov (published by Ya. Ya. Yerzhemskiy, [23]), but in the literature [4] this has come to be known as the Teodorovich hypothesis. V. P. Maslov suggested that the formation of a macrostylolite surface may perhaps be due to the chemical weathering of the solid carbonate buttom, which may have been temporarily dry. V. N. Dominikovskiy [4] asserts that G. I. Teodorovich's hypothesis is superior to the preceding one. V. A. Maslov [10] also explains the origin of stylolite seams in accordance with the views of G. I. Teodorovich.

V. A. Uspenskiy, V. N. Kholodov and K. K. Zelenov, on the other hand, have expressed their opposition to the karst hypothesis. In studying the Cambrian limestones of the Angara region, V. A. Uspenskiy [19] noted that the stylolite seams in them do not always lie along the bedding but sometimes intersect it; this is a circumstance which the hypothesis of solution of a carbonate bottom cannot explain. According to Uspenskiy's analyses, the chemical composition of the fillings in the stylolites and of the insoluble residue in the surrounding rock is the same (in % of insoluble residue):

Rock	SiO_2	Al_2O_3	Fe_2O_3	CaO	MgO	Ignition loss	Carbon
Limestone	70.08	5.12	—	0.36	3.13	18.94	12.00
Interlayer	70.09	8.44		0.24	4.54	15.23	10.93

This similarity in composition indicates that the stylolite filling originated through solution of the limestone or dolomite.

In the stylolite filling from the Lebedev drillhole the content of MgO was as much as 30% along with 4% CO_2, indicating the presence of magnesium silicates. The organic matter in the stylolite fillings has a higher acidity than that in the limestones.

In mentioning the widespread distribution of stylolites in the Cambrian rocks of the Angara region, both in plan and in section, V. A. Uspenskiy casts doubt on the possibility that there could have been numerous repetitions of the exceptional conditions of lack of uniformity in the basin. More likely, in his opinion, the stylolites were produced by some process occurring only once, though over a long time. This can only have been one of the secondary processes taking place after deposition in the fully consolidated limestone. The solution of the carbonate rock that led to the formation of the stylolites may have been considerable, causing a decrease in the thickness of the carbonate strata and even, perhaps, the appearance of brecciated zones.

V. N. Kholodov [20] has written that the Paleogene limestones of Central Asia have been observed to contain stylolite seams that intersect calcite veins, gastropod shells and oölites. When these elements are intersected by stylolites there is a displacement of parts of the veins relative to each other, and part of the material composing the oölites or mollusk shells has disappeared. From these fact he came to the conclusion that the stylolite seams studied by him were formed in the already indurated rock by solution along fractures. V. N. Kholodov's conclusion that stylolites are the result of a peculiar development of tectonic fractures cannot be considered correct, since stylolites also occur in tectonically undisturbed areas.

K. K. Zelenov [5] has observed stylolite seams intersecting calcareous algae (conophytons), which occur in the Proterozoic and Lower Cambrian deposits of Siberia. The G. I. Teodorovich hypothesis, as K. K. Zelenov points out, is unable to explain the branching of a large stylolite seam into smaller ones which again come together in a large seam, the fact that stylolites and stylolite seams intersect each

other, or the intersection of the bedding of the rock and the calcareous skeletons of organisms by stylolites.

New opinions on the origin of stylolites, which resemble the views of G. I. Teodorovich, have recently been published by F. F. Rybakov [12]. To analyze these opinions, it will be necessary to make a small digression. F. F. Rybakov's unusual article begins with a criticism of previous investigators, and with the statement that "their theories were hypothetical, were not based on sufficient factual data and were to a considerable degree haphazard, explaining only the morphological character of stylolites. The widespread occurrence of stylolites... has led many investigators to believe that their distribution is random and chaotic, without relation to any geologic laws" [12, p. 108]. In contrast to previous scientists, F. F. Rybakov stresses that his article contains many new contributions to the problem of the origin of stylolites, and that in it "stylolites are considered to be the result of specific facies and paleogeographic conditions and, finally, as an element of the cyclical nature of sedimentation".

It must be said that the previous investigators (D. V. Nalivkin, M. S. Shvetsov, G. I. Teodorovich, V. N. Kholodov, K. K. Zelenov, A. V. Kopeliovich and others) have also considered stylolites to be the result of definite geologic processes; as regards their role as indicators of certain conditions and their participation in the rhythmicity of sedimentation, F. F. Rybakov cites no material in proof of his assertions. Thus his method amounts to no more than an ill-founded and broadly prophetic declaration.

The new definition of stylolites given by F. F. Rybakov is the following: "Stylolites are straight and curvilinear striated columnar and conical formations of the same rocks in which they occur. In cross section they (which? — the stylolites, or the conical formations of the rock? G. B.) they have the shape of either a curve or a six- or five-faced polygon". "The cross section of a stylolite is a sutural line, which is frequently called a sutural seam" (ibid., p. 108). This definition is false, and the explanation is illiterate.

In the first place, the striations of the stylolites are not random, but strictly regular, and do not intersect; in conical stylolites they run from the apex to the base of the cone, and in prismatic stylolites along the sides of the prism. In the second place, the cones and prisms are not formed only of the same rock in which they occur, but also of other rocks: there are stylolites at the points where some minerals are introduced into others [6], such as marl into limestone, sylvite into halite, and vice versa. In the third place, the cross sections of the "columnar and conical" formations never take the form of a five- or six-sided polygon, but are more complicated. Finally, polygons do not have faces (faces are planes) but sides (lines); it is not clear what Rybakov means by "cross section of stylolites"; it is inaccurate to say that "the cross section of a stylolite is a sutural line", since sutural lines are merely small stylolite seams [22].

"The opinion has been put forth in the literature," writes F. F. Rybakov, "that stylolites are usually perpendicular to the planes of fracture or to the bedding, and that in inclined beds they thus tend to be vertical. The material collected by the author testifies otherwise: stylolites are encountered in the vertical, inclined or horizontal form, but the first predominate and the others are considerably less widespread [12, p. 103]. At first glance it may seem that the author is communicating an important discovery, but after an attentive reading it will be readily seen that this author also finds that stylolites are usually perpendicular to the bedding planes. Thus F. F. Rybakov adds nothing new, and moreover makes a number of errors.

In his conclusion F. F. Rybakov asserts: "It appears that stylolites are not the result of some secondary processes, but are formed during the formation of the sedimentary deposit. A stylolitic horizon is an element of the cycle that characterizes its initial or final stage. Rapidly changing conditions of sedimentation, accompanied by exposure of the sediments to the earth's surface or their raising close to the surface during the formation of the cycle, are evidently the most favorable for the formation of stylolites" [12, p. 111].

Nevertheless the existence of branching and intersecting stylolites contradicts F. F. Rybakov's conclusions. He himself passes silently over the discrepancy between the horizontal and inclined stylolites "discovered" by him and his own conclusions.

Thus the hypothesis of karst formation of stylolites, proposed by G. I. Teodorovich and further developed by F. F. Rybakov, is overthrown by the data cited by V. A. Uspenskiy, V. P. Kholodov, K. K. Zelenov and D. Rigby, as well as the new materials collected by the present writer.

The division of the process of formation of stylolites into three stages, as has recently been done by G. I. Teodorovich [18], does not rescue the situation, since the first stage — the formation of a microkarst — has absolutely no connection, either in its morphology or in the essentials of the phenomenon, with the later stages and bears no relation to them.

Figure 19 shows stylolite seams passing on both sides of a gastropod shell. Part of the shall has been dissolved at the contacts with the stylolite seams. A similar phenomenon is to be

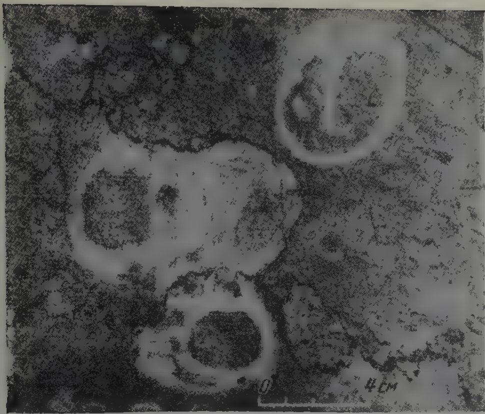

FIGURE 19. Stylolite seams on both sides of a gastropod shell in limestone.

Upper Jurassic, Crimea.

seen in Figure 20, where the stylolite seams pass between the cups of corals. Here also there has been a solution of the material of the coral as a result of stylolitization. It is completely impossible to explain these phenomena by the microkarst hypothesis.

I. G. Chentsov [21] attributes the occurrence of sutures and intersecting fractures to the oxidation and alteration of organic matter, thus saturating the water in the ooze with bicarbonate ions and making it capable of dissolving lime. In his opinion these structures were formed mainly during diagenesis and epigenesis. But the facts he cites, that the stylolites seams and sutural fractures cut through oölites, calcite veins and faunal fragments, testify only to epigenetic stylolitization.

Denticulation, which somewhat resembles stylolites, can be of extremely varied origin. This applies, for example, to the denticulations formed in dirty snow melting under the heat of the sun, or to those formed on the surface of a limestone or dolomite corroded by rainwater (Figure 21), or those resulting from desiccation or other consolidation of sediments. Stylolites differ from these in that they do not form closed structures; along the stylolite seam the parts of the rock become separated, if they are not cemented. The denticular seams in the cracked asphalt pavement of a road differ from stylolite seams in lacking any signs of solution and in the composition of the material filling the cracks. The filling in a stylolite has no relationship to the insoluble residue of an asphalt pavement.

As a result of solution under local pressure, small pits are formed on the surfaces of boulders and pebbles; these have been repeatedly described in the literature. The first to call attention to them was T. Fuchs [24]. He found that a careful study of the contacts between pebbles revealed a denticular surface, like that of the sutures. Similar pitted boulders were observed by V. I. Chalyshev in the Upper Pechora subformation of the Kazanian stage on the Pechora River. The light-colored pits mark the contacts between boulders; these pits frequently contain pressure fractures.

From all the material that has been set forth above, it is possible to arrive at a definition of stylolites and sutures: these are denticular surfaces in a solid rock, formed as a result of solution of the rock under pressure oriented in one direction.

The importance of stylolites is that in the study of sedimentary petrology they are an aid in deciphering the process of epigenesis. For

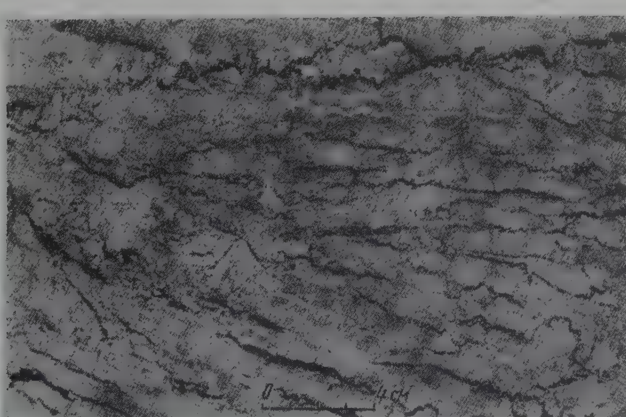

FIGURE 20. Stylolite seams between coral stems.

Acicular stylolites at lower right. Upper Jurassic, Crimea.

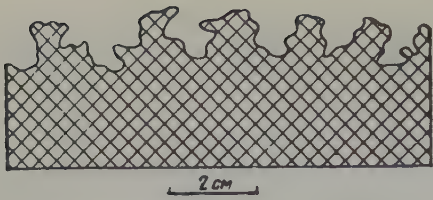

FIGURE 21. Section through a denticular surface in limestone corroded by rainwater -- microkarst.

Travertine from the vicinity of the village of Davalu, Armenia.

example, on the basis of the magnitude of the denticulations, the quantity of material that has disappeared from the structural elements in the rock, whose original dimensions are known, and the amount of clayey matter in the stylolite filling as compared to the general content of such material in the rock, one may form some judgment of the amount of material dissolved during epigenesis. It must also not be forgotten that the rock itself is altered as well, becoming enriched at the expense of the components that are dissolved during stylolitization.

The clayey fillings in stylolites in carbonate rocks may be mistakenly identified, and are taken by some geologists to be interlayers formed through successive layering caused by tectonic oscillatory movements, or other fantastic reasons.

The amount of material removed in stylolitization is anindication of the intensity of the circulation of waters in the rock. Some of the material may be redistributed by the closing of pores during stylolitization.

In some regions stylolites are confined to certain particular layers or strata, thus making it possible to correlate the strata. In so doing, however, one must keep it in mind that stylolite seams may gradually wedge or thin out. As V. N. Kholodov has pointed out, the stylolites that are well developed on the flanks of folds will tend to disappear toward their crests.

From the direction of the stylolite seams and denticulations, one may judge the orientation of the pressure, and from the various generation of stylolites determine the changes in this orientation during the course of tectonic events, or infer these events from the changes in the stylolites. But areas that are highly stylolitized with the removal of considerable masses of rock may give the impression of false unconformities and small folds that are actually the result of downwarping over such areas where material has been carried away. The lines marking the elevation of the stylolite teeth that were formed before the folding are perpendicular to the plane of the beds, whereas those formed after folding are oblique to it.

The result of stylolitization is solidification of the rock, decreasing its porosity and permeability and thus causing a deterioration in its reservoir properties. Permeable sandstones may be rendered impermeable, or even come to resemble magmatic rocks. Uneven stylolitization of carbonate rocks may lead to solidification of some parts and to brecciation and increases in the permeability of other adjacent parts.

Ye. M. Smekhov [14], in studying the reservoir properties of the rocks in fractured oil-bearing strata, found that the highly convoluted, frequently almost denticular fractures in the carbonate rocks resemble sutural lines, and near them the porosity is closed to the greatest degree. The fractures are often filled with bitumen.

Similar observations have been published by M. T. Heald [25]. He determined that the porosity of a sandstone at the distances of 0.5, 1.0, 1.5 and 4.1 cm from the stylolite seam increased, respectively, to 11, 18, 22 and 26%; the permeability of the sandstone increased along with the growth of its porosity. The author states that the presence of clay facilitiates the development of stylolites. The stylolite seams in a porous rock may be the source of a large amount of the cementing material.

In carbonate ore breccias, the ores are enriched through the removal of the carbonates during stylolitization. The presence of stylolites also causes deterioration of the quality of building stones and marbles. The content of the silicate admixtures in average samples of carbonate rocks changes within short distances according to the changes in stylolitization.

I. G. Chentsov [21] has shown that radioactive substances (Figure 22) accumulate in the stylolite seams of certain limestones; these substances include uranium pitch in the form of very small nodular segregations and powdery uranium black, black and blue-black platy vanadium minerals and calcium vanadinates — melanovanadite, hewettite and rarely rauvite.

A. V. Kopeliovich [8] has expressed the opinion that the ore components liberated during the stylolitization of sand grains may form ore-bearing solutions. In this connection, it is necessary to consider two types of stylolitization: 1) stylolitization within a space closed to circulating waters, so that the chemical composition of the rock remains constant, and 2) stylolitization under the conditions of circulating solutions that remove and bring in various substances. In the first type, the dissolved substances are reprecipitated in the pores, in the same places in the rock, taking the form of

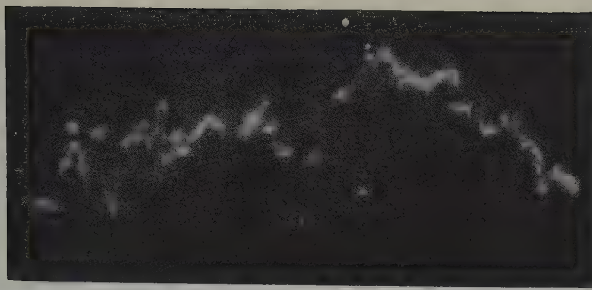

FIGURE 22. Accumulations of organic matter and uranium black (white in photograph) in a stylolite seam.

Radiograph, after I.G. Chentsov, 1959. Natural size.

egeneration borders and new minerals. The econd type must be thought to occur when the mount of dissolved material clearly exceeds at of the newly formed minerals.

Slip planes and striations are probably form- l not only by mechanical polishing and slip- ıge, but also by solution in areas of greatest ressure, similar to the solution in the forma- on of stylolites.

In the theory of metamorphism, in addition the physico-chemical factors one must also ke into account the factor of crystallizational ressure and the corresponding solution of the ırrounding matter under the local pressure us created, as in the growth of stylolites. The agnitude of the crystallizational pressure in um, according to V. Ya. Khaimova-Mal'kova's w measurements, may reach 20 kg/cm^2. Its crease depends on the increase in the con- entration of the solution. It will readily be een that its magnitude is sufficiently great that can be manifested on a large scale in the ourse of geologic time. In particular, within edimentary rocks the replacement of calcite by lomite or quartz, of quartz and carbonates by rite and other such cases is probably to be ex- ained by the solution of the primary solid sub- ance under the local pressure of the growing ystals, when the physico-chemical conditions e favorable.

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THE DEVONIAN BASINS OF THE GORNYY ALTAY AND THE PROBLEM OF THE AGE OF THEIR STRUCTURES[1]

by

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This article presents a brief review of the data on the development of the basins existing within he Gornyy Altay in Devonian times, among which there were some extremely active basins of the ype of flysch troughs. After examining the data, the author comes to the conclusion that at the eginning of the Hercynian tectonic epoch, when the Ob'-Zaysan geosyncline was formed along the vestern margin of the Altay-Sayan region, the Altay to some degree served as its limiting frame r boundary, though it was itself far from stable. The Gornyy Altay, in particular, was a mobile one of folding with individual still active or newly appearing basins of high mobility (some of these nay be considered as short-lived lateral branches of the Ob'-Zaysan geosynclinal system).

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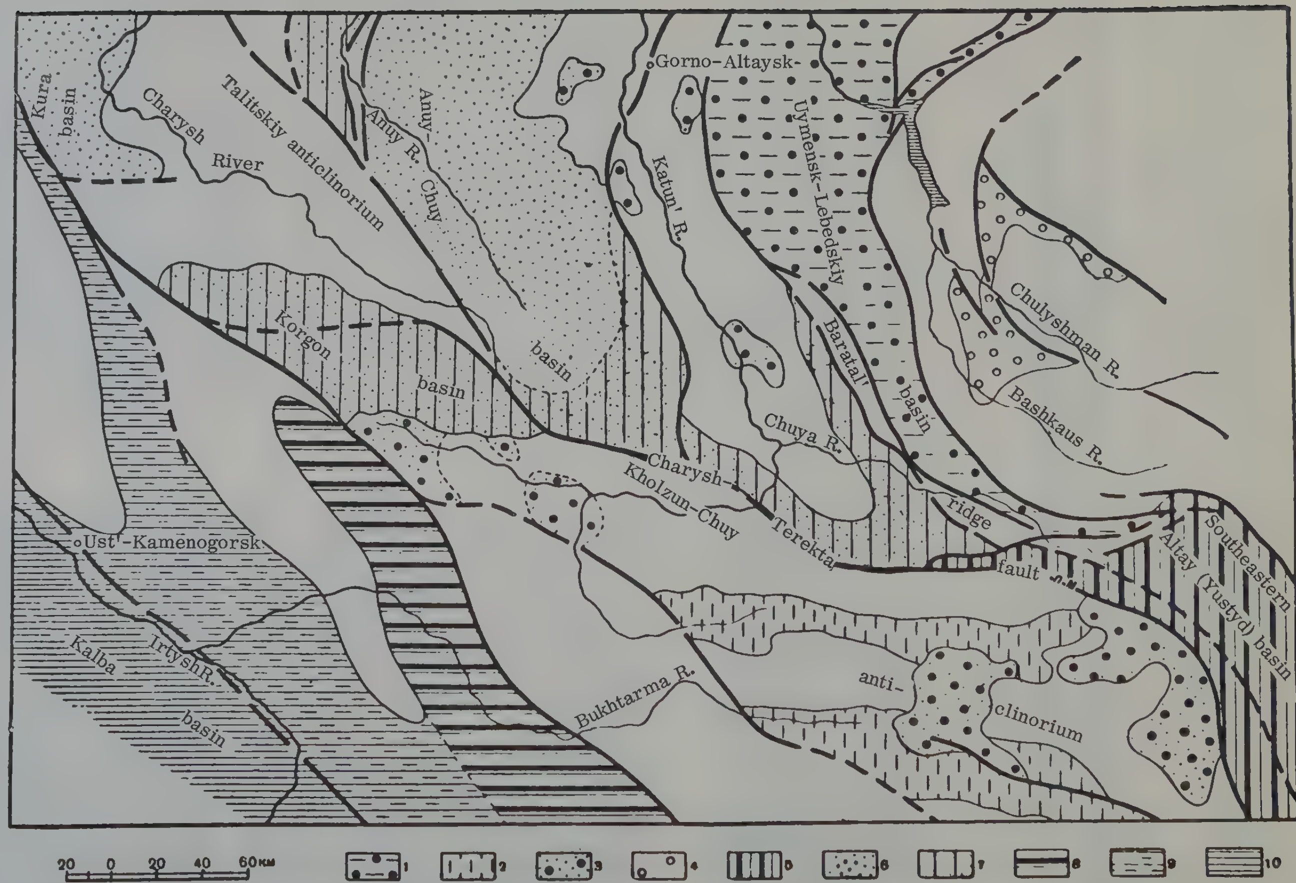
In the "Bulletin of the Academy of Sciences f the U. S. S. R., Geologic Series", No. 11, 958 there was published an article by A. L. Matveyevskaya entitled "On the Position of the Rudnyy Altay in the Structural Plan of the ayan-Altay Region" [13]. The author of that rticle cites some very interesting material howing the common character of the geologic evelopment of the Kolyvan-Tomsk and the rtysh-Zaysan (Kalba) geosynclinal zones, vhich she consideres to be the separate branches f a single Ob'-Zaysan Variscian geosyncline; he structures of the Rudnyy Altay she believes o be of more complex origin. Some of the atter are blocks of older — Caledonian — structures uplifted or downwarped during the period f the Variscian tectonic movements (in the atter case, that of their lowering, they are imilar to the marginal basins of the Kolyvan-Tomsk zone); others, in Matveyevskaya's pinion, are related to the younger — Variscian — geanticlinal uplifts of the Aleyskiy anticline.

In setting forth these conclusions, A. L. Matveyevskaya disputes those of D. I. Gorzhevskiy, who in one of his papers [6] unjustifiably compared the Gornyy Altay to the Kalba geosynclinal zone in the degree of its mobility luring the Middle Paleozoic and contrasted both hese regions with the Rudnyy Altay, which contains blocks of ancient (pre-Variscian) structures. During the course of this controversy Matveyevskaya, for reasons not clear to the reader, directed the chief burden of her critical remarks not against the aforementioned paper by D. I. Gorzhevskiy [6], but against one of the present writer's publications [2]. The impression was created that it was this writer's data, in particular, that led to the erroneous conclusions regarding the age of the geologic structures of the Gornyy Altay. For this reason this writer considers it his duty to elucidate the conceptions he had evolved on this complicated problem and to introduce some necessary clarifications based on new materials. At the same time he will be forced to make certain criticisms of the article by A. L. Matveyevskaya [13].

One of the principal ideas expressed by this article is that the Rudnyy Altay is situated at the zone of contact between the Variscian geosynclinal system and the epi-Caledonian platform of which the Gornyy Altay is a part (in characterizing its structures, such terms as synclise and autoclise are sometimes used). It is, however, unfortunately impossible to agree with this conclusion, especially its second part. Even without regard to the fact that the division of the Paleozoic structures of the Altay-Sayan region into Caledonian and Hercynian is a highly tentative one, it is still impermissible to use such a term as "young platform" for the Gornyy Altay during the period of time under consideration.

Was the Gornyy Altay at that time actually a young platform? The most important feature of this region as a marginal area of the Caledonian structure of Siberia lies precisely in the fact that throughout the entire Middle Paleozoic (through the Early Carboniferous) the Gornyy

[1]Devonskiye progiby Gornogo Altaya i vopros o vozraste yego struktur, (pp. 58-66).



Sketch map showing the disposition of the Devonian basins of subsidence within the Gornyy Altay:

1 - long-developing (from the Ordovician to the end of the Devonian) secondary geosynclinal basins of subsidence upon an Early Caledonian folded basement reactivized at the beginning of the Givetian age; 2 - briefly existing (from the Silurian to the Eifellian age) interior basins upon later Caledonian anticlinoria; 3 - superimposed basins of the first half of the Devonian (Early Devonian and Eifellian age); 4 - superimposed basins of the second half of the Devonian (Givetian age and Late Devonian); 5 - flysch basins of the second half of the Devonian, developed along deep faults in the zones of pre-Devonian (possibly pre-Silurian) fold folding; 6 - residual geosynclinal basins terminating during the Devonian or in post-Devonian times; 7 - zones of transition from the residual basins to marginal and superimposed basins, with a considerable development near faults of black shale flysch series and with intensive volcanic activity in the Eifellian age; 8 - long-developing (from the Early Devonian to the Early Carboniferous) flysch basins of subsidence; 9 - Hercynian basins of the Rudnyy Altay type (with considerable mobility and intensive igneous activity); 10 - Hercynian interior geosynclinal basins.

Altay was without any possibility of doubt a mobile folded zone. Moreover within its area at that time there still continued to be places of highly mobile subsidence. In other words, if the Rudnyy Altay was a region of transition and mutual interpenetration of Caledonian and Hercynian structures, but tending more toward the Hercynian, the greater part of the Gornyy Altay may also be related to this transitional zone, but with a greater predominance of Caledonian structures. In the Rudnyy (or more generally the Southwestern) Altay the Caledonian structural elements are fragments of older structures existing among predominantly younger ones, whereas in the Gornyy Altay, on the other hand, the Hercynian structural elements penetrate the generally Caledonian area as separate still active basins of subsidence and mobile zones.

Let us turn to the facts. In the southeastern part of the Gornyy Altay, in the area of the Chuya valley, there are thick deposits of the flysch type belonging to the second half of the Givetian stage and the Upper Devonian. These are combines as the Yustyd series, which includes the following three formations: 1) the Tashanta formation of the Upper Givetian substage, with Chielevian fauna in its upper part; 2) the Bar-Burgaza formation of the Frasnian stage, with Upper Devonian fish fauna in its lower, multicolored strata and Anathyris phalaena Peetz higher up in the section; and 3) the Boguta formation, tentative assigned to the Famennian stage, with a terrestrial so-called "Archaeopteris" Upper Devonian flora. In the cyclical character of the predominant rocks, in the hieroglyphs at the base of each cycle and in many other lithological peculiarities, these deposits more than any other in the Altay resemble the classic flysch (strata with marine or fresh-water fauna and packets of multicolored rocks occupying a subordinate position among the main deposits). In essence, these are the same black-shale flyschoidal strata which A. L. Matveyevskaya with full reason considers typical of the Variscian Ob'-Zaysan geosyncline (passing over, of course, their development in "synclises of the epi-Caledonian platform"). Their thickness in the Gornyy Altay is as much as 6000 m. These deposits occur over a relatively narrow zone along an enormous "axial" deep fault (or group of closely adjoining faults) in the Southeastern Altay. These faults branch out westward and northwestward to form, at the western margin of the Chuya valley, an entire fan of regional faults, including such well known faults in the literature as the Kurayskiy and the Charysh-Terekta (Bashchelak-Southern Chuya) faults. Toward the southeast the "axial" fault (group of faults) continues into Mongolia and gradually bends southward. This same trend is followed by the zone of black-shale deposits that fill the adjacent basin of subsidence. Farther to the south, within Northwestern Mongolia, this basin evidently comes to a blind end.

Formations of Early Devonian and Eifellian age, exposed along the margins of the basin and in the adjoining anticlinoria, form patches of isometric form which are clearly of a superimposed nature, and are represented by rocks typical of the late stages of development of the mobile zones of the earth's crust. These coarse clastic red beds and volcanogenic continental, sometimes gypsiferous, strata (the Ulandryk and Aksay formations) have a total thickness of some 3 km or more. Within the Kholzun-Chuya anticlinorium — south of the basin — the superimposed graben-synclines contain a Silurian stratum of small thickness (which may have been accumulated only within narrow interior basins formed by pre-Silurian movements along longitudinal faults in the crest of the anticlinorium). This has little in common with the typically marine Silurian geosynclinal deposits of the Anuy-Chuya synclinorium that forms the bottom of the upper structural stage; the lower structural stage of the anticlinorium is represented by Early Paleozoic rocks that were tightly folded in pre-Arenigian and pre-Caradocian times. The deposits of the upper half of the Devonian overlie the Silurian without any visible angular unconformity, but crop out outside the aforementioned graben-synclines to form the isometric patches of outcrops pointed out above.

Thus we have here the successive stages in the extinction of a geosyncline and its transformation into a folded zone during the Ordovician, Silurian and first half of the Devonian periods. But in the second half of the Devonian, as a result of the violent tectonic movements at the end of the Eifellian and beginning of the Givetian stages (or at the beginning of the latter), the picture changes strikingly. Within the zone corresponding to the major tectonic seam the mobility of the earth's crust again rose sharply; this led to the final formation of this basin of subsidence. The fact that this basin was indeed relatively narrow (some 40 to 50 km in width) and thus scarcely exceeded the limits of the present zone of Upper Givetian and Upper Devonian deposits, is indicated by the facies changes at its margins (the more coarsely clastic or more multi-colored facies lying next to the tectonic faults that border the zone).

On the basis of its outlines and its geographical disposition, as well as its tectonic regimen during the time of deposition of the black-shale strata, this basin may be placed in the category of highly mobile basins of the type of flysch troughs (and it is well known that flysch troughs do not form in the interiors of platforms, even young ones!). It has nothing, of course, in common with synclises. It is necessary to stress the independence of the basins if the type considered here — that is, the need to place them in a special category of their own. In this particular case we are dealing with a mobile zone connected to an ancient tectonic

seam: in essence, a rejuvenated basin of subsidence which has already been transformed by previous phases into a folded zone, whose reactivation is due to new and more violent movements that are in turn the echoes of the birth of new geosynclines in the neighboring regions. These movements, which were sufficiently violent to break up the not very strong "solder" healing the seam, in all probability led to a considerable reworking of the sunken basement, but only within a certain narrow block (wedge) corresponding to the zone of the deep fault (or group of faults). This supposition, the present writer believes, most convincingly explains the burst of tectonic activity immediately before "death", concentrated in the narrow flysch troughs, which occurred on the eve of the total extinction of the geosynclinal system, when half or more of its territory had already been turned into a folded belt.

However this may be, without going into any further consideration of the nature of flysch basins in general, it may be said that the data obtained in recent years point to exactly such a development in the above-described basin of subsidence in the Southeastern Altay.[2]

Let us now touch upon some other basins of the Gornyy Altay, which maintained a considerable mobility during Devonian times.

The faults that are concentrated together into a single parallel group in the area of the Chuya valley, as already mentioned, branch out into a broad fan farther to the west. Between two of these, the two most northeastern faults (running along the southern slope of the Kuray Range and thence through a part of the Aktash, into the area of the left bank of the Bashkaus River, the Sumul'tinskiy Range and farther northward), lies the so-called Uymensk-Lebedskiy basin, which was superimposed upon the Early Caledonian—Cambrian folded basement and acted as a basin of sedimentation for a very long time: from the Ordovician, apparently, to the end of the Devonian. Characteristically, it underwent a considerable rejuvenation after the Middle Devonian stage of folding. Only during the Givetian age were sediments totaling 2.5 to 3 km in thickness accumulated here; these are represented primarily by coarse clastic continental deposits and volcanogenic rocks (most prominently porphyritic tuff conglomerates). Still one more such thick sedimentary-volcanogenic series was deposited at the beginning of the Late Devonian (after the Chielev transgression). The southeastern, highly compressed end of this basin in the zone of the northern margin of the Chuya valley, at Mt. Taboshak, was apparently connected with the above-described basin of the flysch type. Northeast of both of these, in the second half of the Devonian there was an extensive area of erosion which served as a source of clastic material for these basins. Southwest of the Uymensk-Lebedskiy basin lay the periodically uplifted Baratal'skaya Ridge, which acted as a cordillera for the western part of the flysch trough (the eastern part of this ridge, partially covered with Devonian deposits, is characterized by coarser clastic and even multi-colored facies, by smaller thicknesses and even by the wedging out of individual strata).

Another pair of faults, which diverge west of the Chuya valley toward the west and the northwest, surround that part of the well known Anuy-Chuya basin in which the Silurian deposits are represented by thick geosynclinal sediments. It may also be considered that the area where these faults merge into each other is also the terminus of the Silurian geosynclinal basin, which was turned into a mobile folded zone toward the beginning of the Devonian. Near the faults that outline this area there are abundant Middle Devonian (probably Eifellian) extrusives, where as the deposits of the second half of the Devonian, of the same age as the black-shale strata in the Southeastern Altay, are represented by somewhat different facies and are, apparently, of much smaller thickness.

Approximately the same type of section of Devonian deposits occurs in the central part of the Anuy-Chuya basin, in the area of the so-called Kuratinskiy graben, where from the main (or Charysh-Terekta) fault there is still one more branch running almost north-south in direction. Here the Lower Devonian is represented by thick multi-colored deposits, probably for the most part continental (the Karakudyurskaya formation).

Devonian deposits similar in nature to those just described also occur farther to the west and northwest, in the area of the Korgon and Tigerets Ranges, but here Lower Devonian deposits are, as a rule, absent. Characteristically, the extensively developed deposits of the Eifellian stage show close similarity to the contemporaneous rocks of the Rudnyy Altay (the Talovskaya formation) and have a great thickness, on the order of 2 to 3 km. Northward they overlie eroded Silurian and Ordovician rocks that were crumpled into folds by the pre-Devonian tectonic phase; in the south they directly overlie the ancient metamorphic schists of the greatly uplifted horst on the northern margin of the Kholzun-Chuya anticlinorium. Possibly the basin that was developed here in Eifellian times did not coincide in space with the earlier existing basin and was destroyed by the pre-Devonian folding, displaced considerably to the south relative to the basin, was again superimposed in connection with the movements along the fault

[2]Another similar basin of subsidence, parallel to to the Eastern Altay but probably of smaller size, and containing black shale deposits of the same age, occurs somewhat farther to the northeast, within Southwestern Tuva and the adjoining areas of Mongolia.

that separated the Kholzun-Chuya anticlinorium from the structures of the Northwestern Altay. At present this fault is to a considerable degree hidden beneath Devonian deposits (perhaps the closing of the fault was facilitated by volcanic activity).

The Middle Devonian — pre-Givetian or Early Givetian — tectonic phase is correspondingly reflected in the Korgon basin.

North of the line connecting the above-described basins, the nature of the Devonian deposits gradually changes toward continually more stable marine and littoral-marine facies in the northern part of the main — Anuy-Chuya — basin. In addition, the multi-colored Lower Devonian sediments are replaced by gray sands and shales or by carbonate-sand-shale deposits; the amount of extrusives of the Middle Devonian is generally smaller (though they are extensive near the faults); Eifellian limestone strata make their appearance and, along with the Famennian deposits, there are strata that grade into the Tournaisian stage of the Lower Carboniferous. According to A. B. Gintsinger's data, the extent of dislocation of the Lower Devonian rocks is approximately the same as that of the Silurian strata. As a result of the superimposition of several tectonic phases upon each other the more intensive ("complete") folding observable in the lower part of the Middle Paleozoic section is gradually replaced upward by relatively gentle "incomplete" folding. This is a typical feature of the development of the basins of subsidence in the northwestern part of the Gornyy Altay, showing their relationship to those of the Salair Range, reflected in the many phases of tectonism with a gradual decline of the geosyncline which, however, did not come to an end at the beginning of the Devonian.[3]

Most likely the final formation of the folded zones on the sites of these basins took place in the Carboniferous. It is this very fact that serves as the present writer's reason for speaking of the residual character of the Anuy-Chuya basin during the Devonian — the period in which the Hercynian geosynclinal systems began their development [2].

[3]A smilar multi-phase character has also been noted in the zone of anticlinal uplifts in the Northwestern Altay, where, according to Yu. A. Perfil'yev, M. N. Bartseva 1 and others, one may at the present time distinguish the follosing major tectonic phases: at the top of the Middle Cambrian, pre-Arenigian, pre-Caradocian, pre-Silurian, boundary between the Lower and Upper Silurian and post-Silurian. The formational nature of the sedimentary deposition between these phases is also very similar to that which has been observed in the Salair Ridge. In general, it may be said that there was a considerable degree of unity in the geologic development of the Northwestern Altay and the Salair. The main difference between them lies in the intrusive igneous activity, which is absent in the Salair and very intensive within the Gornyy Altay (Talitskiy massif).

Evidently there were somewhat different tectonic conditions during the Devonian in the area of the village of Kur'ya in the northwest of the Altay (approximately in the axial part of the long-developing synclinorium that lay between the Talitskiy and the western end of the Kholzun-Chuya anticlinoria). In the carbonate facies of the upper strata of the Lower Devonian and the Eifellian stage (the Murzinskaya formation), as in the entire history of its geosynclinal development, this area (even more than the Anuy-Chuya basin) recalls the Salair and deserves the name of a residual basin. But as early as the second half of the Givetian stage it ceased, in essence, to be a basin of subsidence. The extrusives developed here, which some authors would tentatively assign to the upper part of the Givetian stage or to the Upper Devonian, overlie the deposits beneath with a sharp angular unconformity (the area of their distribution being considerably displaced to the side relative to the older Devonian rocks), whereas the Tournaisian marine deposits occur as isolated exposures within a field of Lower Paleozoic rocks.

From this brief review of the Devonian basins it will be seen that at the beginning of the Hercynian tectonic epoch — throughout the whole of the Devonian — the Gornyy Altay was generally still in the stage of a mobile folded belt. At the margins of the positive structure; as in the basins which had lived through the main phase of folding, formations typical of the late stage of geosynclinal development were produced in the first half of the Devonian period. In the second half of the Devonian some of these areas were eroded or became the sites of brief sedimentary deposition at the maximal — Late Givetian — transgression of the sea. In other similar places during this time, on the other hand, there was an intensification of downward movements, leading to abundant accumulation of deposits of approximately the same type (that is, characteristic of the late stages). Along with this, within the Gornyy Altay there were also still open (residual) geosynclinal basins, where through the whole Devonian there continued to be a fairly intensive accumulation of sedin ents resembling partly those of the middle and partly those of the late stages of development, as well as basins of especially great mobility intensively developing along zones of major faults during the second half of the Devonian, when relatively fine-grained flysch sediments up to 6 km thick were accumulated here. Both these kinds of basins soon, apparently at the beginning of the Carboniferous, ended their existence.

Very high mobility, in addition to the above-described flysch troughs, also characterized the peculiar secondary basins that developed on the Early Caledonian folded basement (such as the Uymen-Lebedskiy basin). After being superimposed as early as the Ordovician as a result of block displacements along large faults, they were again renewed in the Devonian, especially after the movements in the middle of the Devonian,

when high mountain ranges appeared on their margins and acted as sources of coarse clastic material; magma rose along the faults to produce lava for the abundant surface eruptions. In the main features of their geological development these basins belong to the same category as the Minusinsk group, but in their tectonic activity in the Devonian, their rate of sedimentaion and their narrowness (especially in the south) they are unique and deserve to be compared to "continental geosynclines". The reason for this probably lies in the same renewal of mobility of the earth's crust along closely spaced major faults, which one is inclined to explain as the superimposition of flysch basins. It is not a matter of chance that the inferred southern end of the Uymensk-Lebedskiy basin merges directly into the basin of the Southeastern Altay and has the same features of a flysch trough, while the Lower Paleozoic strata of the so-called zone of Kurayskiye schists and gneisses that crop out from beneath the Devonian mantle deposits show the signs of intensive metamorphism and granitization at depth. They may be considered to be the reworked basement of the basin that was superimposed in the Ordovician and renewed again in the Devonian; in this repeated renewal it is possible that the wedges of Ordovician rocks were drawn into the zone of metamorphism at depth.

It should be noted that the morphology of the structures in the folded zones that arose on the site of the most mobile Devonian basins of the Gornyy Altay is, in the majority of cases, quite peculiar and testifies to the intensity of the folding. In places the folds are of the highly elongated "linear" type, with very great amplitude; there is regional overturning of the folds, which are sometimes isoclinal, etc. In places one may even see folds of the same type as those produced experimentally by N. B. Lebedeva in her soft plastic material by means of a vertical upward movement of a narrow rigid plunger. An example is the Kyzyl-Shinskiy anticline in the Southeastern Altay.

The final stages in the development of the Gornyy Altay, which marked its transition from the state of a mobile folded zone to that of a young platform, took place in the beginning or middle of the Late Paleozoic (the Middle to Late Carboniferous or the Permo-Carboniferous). At this time were formed the continental coal-bearing strata that fill the small intermontane basins — grabens — in the eastern part of the region, at the southern and southeastern continuation of the Kuznets basin.

How is one to reconcile all that has been said here with the view which A. L. Matveyevskaya has, in this writer's opinion quite correctly, stated, that the Ob'-Zaysan Hercynian geosyncline is a younger structural zone than the entire Altay as a whole (both the Rudnyy and the Gornyy), where there were earlier formed Caledonian structures?

The answer to this question must evidently be sought in the fact that a folded region (in this case, the Altay) may to some degree serve as a limiting frame or boundary for an adjacent younger geosynclinal system beginning its development, even before the older one has become completely stabilized. In such a case there is a partial overlapping of the two adjacent (in time) tectonic epochs, such as the Caledonian and Hercynian. The same tectonic phase may in one area be one of the closing stages of the earlier epoch of folding, whereas in the other (adjacent) area it may be one of the early phases of the new epoch. The residual (still active) basins that may be discerned within the older region can be considered as lateral branches of the younger geosyncline, but the time of their existence is, within the limits of the new tectonic epoch, considerably shorter. The individual tectonic phases that may be distinguished by their intensity, which formed the young geosyncline, were capable of causing a partial renewal of the older one, as reflected in the sporadic transition of some zones from the later back to the middle stages of development (possibly accompanied by anatexis of their "root"). Nevertheless such a burst of tectonic activity, concentrated in limited areas, could not continue for long, and by the time the nearby younger geosyncline has gone through the period of its most intensive existence, the older basins had ceased completely to be active.

The examination of these problems convinces the present writer that in studying the tectonic history of any region one must keep in mind the possibility of a considerable retardation of the later and terminal stages of its development, during which for several geologic epochs the region in question is in a state of transition, combining features of a young platform with traits of a mobile folded zone (incompletely ended or renewed basins, which are far from platforms in their nature). This illustrates the error of a formal approach to the classification of the structures of such regions, using such terms as synclise, anticlise and marginal basin without qualification. For example, although the structures of the Rudnyy Altay may with some justification be considered those of marginal basins, it must be stressed that these are not at all the basins that usually occur on the periphery of fully formed platforms (the pericratonal zone with intensive igneous activity).

Thus the Gornyy Altay, in the features of its development and in its effect on the structure of the adjacent regions during the period of the Variscian tectonic movements, displayed very few traits of a young platform. One may justifiably speak of platform conditions here only during Mesozoic times, when the entire Altay-Sayan region as a whole (tentatively distinguishing Hercynian from Caledonian structures within it) had fully terminated its development as part of a single extensive zone of Paleozoic folding. Since this conclusion has recently and with

sufficient clarity been set forth in a paper authored collectively by a group of geologists of the All-Union Aerogeological Trust [9], there is no need to dwell upon it in detail.

In conclusion, the writer wishes, on the basis of the materials on the Altay, to draw the attention of specialists in flysch deposits to the problem of the historical position of this formation in the successive series of geosynclinal formations. The flyschoidal sand-shale strata, which are less definite in their features, here occupy the beginning of the middle stages of development, as indicated in the stratigraphic scheme by Yu. A. Bilibin and others (Cambro-Ordovician and Ordovician deposits). But the flysch, in the narrower sense of the word, in the Altay usually appears after the creation of certain formations typical of the late stages of development. They are deposited in connection with the secondary increase in the mobility of the earth's crust in zones of deep faults, during the burst of tectonic activity immediately before the "death" of the declining geosynclinal system.

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THE USPENSKIY ZONE OF CENTRAL KAZAKHSTAN AND SOME OF ITS ANALOGUES[1]

by

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The Uspenskiy zone is a relatively narrow belt of intensively deformed Devonian and Carboniferous volcanogenic and sedimentary rocks, cut by granite intrusions, stretching for many tens of kilometers from west-southwest to east-northeast, from the village of Atasuyskiy to Mt. Karkaralinsk. In the north this zone abuts on the Tekturmass zone; in the south it adjoins the Ortaushet-Shoka uplift, composed of older formations.

Different opinions have been put forth at different time in regard to the tectonic structure of the Uspenskiy zone. First a zone of tight folding was distinguished, bounded on the north by a single main structural seam; south of this seam the maps showed compressed rocks with intensive cleavage, and north of it a series of younger rocks [21]. This zone of tight folding and crumpling was then identified as a thrust zone, inclined southward and having a vertical amplitude of no less than 1 km, and more than 2 km along the dip [9].

Later the territory of this zone was somewhat extended and was characterized as the Uspenskiy synclinorium, whose flanks were composed of Lower and Middle Devonian extrusives and whose upper parts were of Upper Devonian limestones and Lower Carboniferous sedimentary-tuffaceous deposits. It has been shown that within the synclinorium the rocks were crumpled into sharply compressed folds that were overturned toward the north; on the southern flank they were broken by cleavage fractures and small disjunctive dislocations; the northern flank of the synclinorium was destroyed by the large Mt. Kaldyrma granite intrusion [3].

Finally, in a number of later works the Uspenskiy zone appears with various characterizations — as a zone of subsidence and development of linear synclinal folds [4], a zone of faults and complex synclinal structures [10], a zone of deep faults which, together with the Spasskiy zone, border the southward leaning ultrabasite layer [6, 15], as a deep mobile zone marked by an intrusive igneous complex [30, 31], etc.

All these interpretations testify to the complexity of the structure and to the important position of the Uspenskiy zone in the general tectonic plan of Central Kazakhstan. They also indicate that there is no unity of opinion on these problems and that the specific tectonic structure of the Uspenskiy zone has still not been determined with sufficient definiteness; its boundaries, the mechanism of its formation, its geologic history and other important aspects are still not well enough known.

The present article, based on materials collected personally by the author during three years' investigations in the field (1956-1958) as well as on other investigators' data, will attempt to solve the problems mentioned just above.

I. SUBDIVISION OF THE STRATIGRAPHIC SECTION INTO FORMATIONS AND SOME RULES CHARACTERIZING THEIR CHANGES

The natural major subdivisions of the Uspenskiy zone and the regions adjoining it, characterized by stable rock composition and paragenetic associations, may be grouped into five formations (from bottom to top): 1) a chert-jasper-diabase formation, 2) a sandstone-shale formation, 3) a lower volcanogenic formation, 4) a sandstone-carbonate formation and 5) an upper volcanogenic formation (Figure 1).

The chert-jasper-diabase formation embraces an assemblage of jaspers, jasper-quartzites, siliceous schists, diabases and diabase-porphyrites, among which there are packets of siliceous siltstones, greenish-gray sandstones, sometimes acidic extrusives and very rare limestones. This includes mainly the Urtyh-Dzhal' series of A. A. Bogdanov [3] (so that the name Urtyn-Dzhal has been suggested for this formation), as well as the Baydau-Letovskaya and Dzhaman-Shurukskaya suites

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[1]Uspenskaya zona Tsentral'nogo Kazakhstana i yeye nekotoryye analogi, (pp. 67-82).

and their analogues. The thickness of these formation totals several kilometers.

The age of these rocks has been variously determined. In the Atasuy area [4] they are assigned to the Riphean (the Urtyn-Dzhal'skaya series) and to the Cambrian (the Baydau-Letovskaya formation); in the Taldy-Espetau Mountains to the Lower Paleozoic or Ordovician [28, 29 et al.]; in the Shetskiy area (southeast of Agadyr) a similar series of sandstones, shales, diabase porphyrites, jasper-quartzites and reef limestones contains a Wenlockian-Ludlovian fauna in its upper part [23]; in the Northern Balkhash area the age of the formation that underlies the faunally characterized

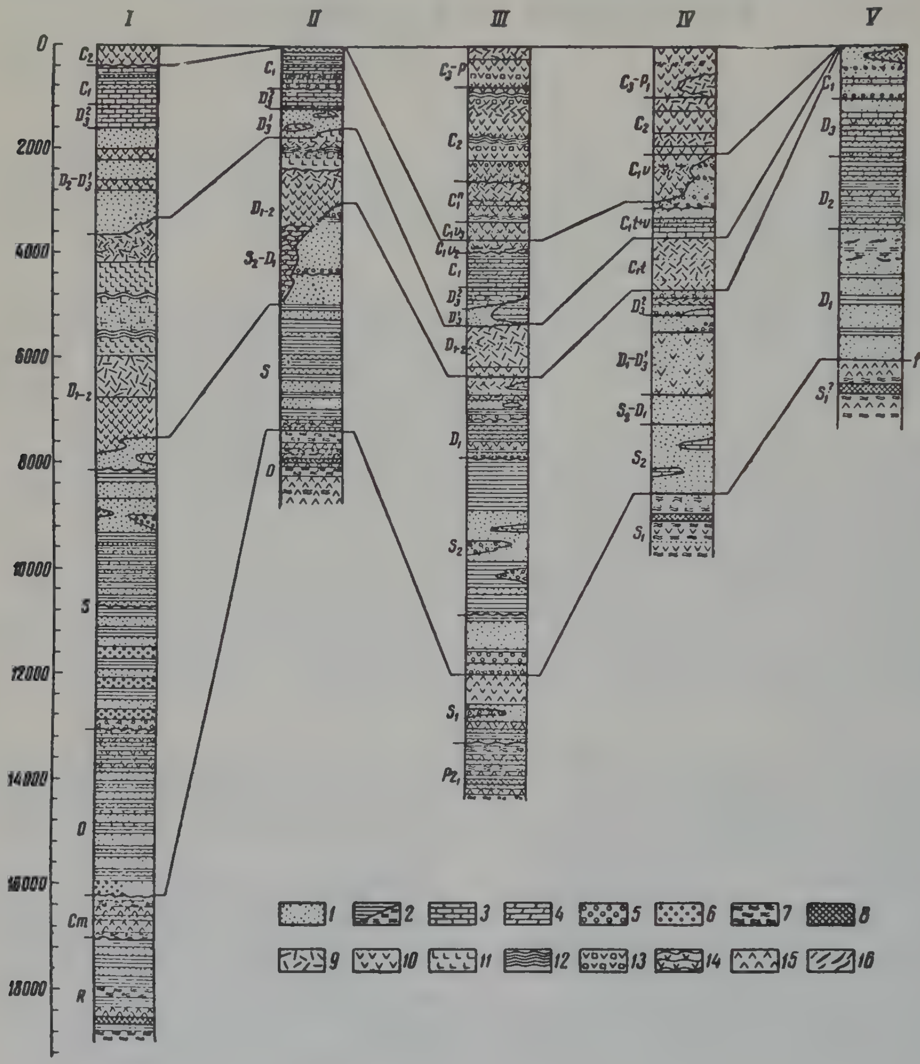

FIGURE 1. Diagram showing the interrelationships of the sedimentary and volcanogenic formations in the section through the Atasu-Galkhash region.

I - Atasuy region (after A.A. Bogdanov, O.A. Mazarovich, A.Ye. Mikhaylov, N.P. Chetverikova and G.I. Bedrov); II - Zhaksy-Tagaly and Taldy-Espetau region (after S.Ye. Kolotuzhina, N.A. Shtreys and A.I. Suvorov); III - Shetsk and Aktogay regions (after G.I. Bedrov, K.T. Kulikovskiy, P.P. Chuyenko, N.A. Pupyshev, M.I. Aleksandrova, and I.I. Radcheko); IV - Northern Balkhash region (after V.A. Vakhrameyev, V.Ye. Luy, Yu.V. Nikishov); V - Eastern Balkhash region (after V.V. Galitskiy, S.N. Golyshev and B.Ya. Ponomarev).
Legend: 1 - sandstones; 2 - argillites and clay shales; 3 - limestones; 4 - marls; 5 - conglomerates; 6 - gravelites; 7 - siliceous shales and tuffites; 8 - jasper-like rocks; 9 - quartz porphyries; 10 - porphyrites; 11 - albitophyres; 12 - lavas; 13 - tuff breccias and tuff agglomerates; 14 - porphyroidal rocks; 15 - diabases; 16 - carbonaceous materials.

Gotlandian rocks is considered by some authors to be Cambrian and by others (the more likely supposition) Ordovician.

The sandstone-shale formation was initially identified and described under the name of the Kayraktinskaya formation [17]. This consists of gray-green sandstones, quartz-chlorite and sericite schists, siltstones, clay shales, argillites and in places limestones and conglomerates. In some parts the sandstones are predominant; in other parts the coarser rocks or the shales. But the general monotonously greenish-gray appearance of the formation and its position between the Urtyn-Dzhal'skaya and the Lower Volcanogenic formations remain unchanged.

In the northwestern part of the territory under consideration, in the Nura synclinorium, the thickness of the Kayraktinskaya sand-shale formation is 8 km, its lower part containing Bryozoans of Ordovician appearance, with a rich fauna of Wenlockian and Ludlovian corals [4] higher in the section. In the Shetsk area its thickness does not exceed 5. 5 km. The lower strata are Wenlockian-Ludlovian in age [23], and the upper, according to their brachiopods and tabulates, are Upper Ludlovian. The lists of fauna also include Devonian species; in the uppermost part of the sand-shale series in the Shetsk area, where it contains small interlayers on extrusives, there is a Devonian flora [1]. Still farther east, in the Northern Balkhash area, the formation belongs to the Siluro-Devonian and was at one time identified by V. Ya. Lui and V. Yu. Nikishov as a "Siluro-Devonian transitional formation" with a definite faunal assemblage. Other authors, supported by detailed paleontological investigations, have also come to similar conclusions [16]. The complex of sand-shale deposits in the Northern Balkhash area, which overlies the jasper-quartzites, extends upward in the section through the Upper Devonian inclusive, containing abundant pure extrusives only in the Famennian Here the thickness of the formation reaches 4 km.

The lower volcanogenic formation is composed primarily of porphyrites, quartz porphyries and albitophyres, along with their tuffs and lava breccias. In places they contain sandstone wedges, which are not, however, extensive. The chief area of distribution of all these rocks, where their thickness exceeds 4 km, is the Sarysu-Teniz uplift; accordingly this formation may be called the Sarysu-Teniz formation. Its age in these places is established by its stratigraphic position between the Silurian deposits and the Middle Devonian deposits of the Frasnian stage [4].

In the Zhaksy-Tagaly Mountains, according to the present writer's data, the extrusives of the Sarysu-Teniz formation are divided into two suites. The lower suite is formed of porphyroid rocks produced from acidic extrusives and their tuffs and tuff lavas; they lie conformably upon the Silurian sand-shale deposits and are themselves of Silurian-Devonian age. The upper suite, consisting of porphyrites, quartz porphyries, albitophyres and felsites and separated from the porphyroids by an erosional surface and an unconformity, is Middle Devonian, so that its upper part may correspond to the Frasnian stage. The total thickness of the lower volcanogenic formation in the Zhaksy-Tagaly Mountains reaches 3 km; in the Shetsk area to the southeast it is no more than 1 km [1]. In the Northern Balkhash area the first interlayers of acidic extrusives of the Sarysu-Teniz formation appear in the Famennian, and in the Tournaisian they form blanket flows from 1 to 2 km thick and are, in turn, covered by sandstone-carbonate deposits with Tournaisian and Visean fauna [7].

The sandstone-carbonate formation includes (from bottom to top): red terrigenous deposits, marine carbonate and terrigenous coal-bearing deposits. The carbonate deposits are silicified and rich in manganese oxides; I. S. Yakovkin has distinguished them as the Uspenskiy series. In a number of places, however, they are closely associated with both the red beds and the carbonaceous deposits and together with the latter form essentially a single complex, which separates the lower from the upper volcanogenic formations. Since all these rocks are found in the Uspenskiy zone, the name Uspenskiy formation is reserved for all of them together.

In the Alasuy area the formation is represented by red-colored and carbonate deposits, and has a thickness on the order of 3 km. Its lower part is correlated with the deposits of the Middle Devonian — the Frasnian stage — while the strata above contain Famennian and Tournaisian fauna [4].

The formation has a similar facies character in the Karayktinskiy area. Here it begins with the Frasnian stage and does not go beyond the Tournaisian; its thickness is no more than 1. 8 km. Still farther southeast, in the Shetsk retion, the upper boundary of the Uspenskiy formation grades into the Visean, and the deposits contain carbonaceous strata with Middle Visean flora; here the thickness has been determined as 1. 6 km [1]. In the Northern Balkhash region the formation is composed of carbonate and terrigenous coal-bearing sediments with total thickness ranging from 0. 6 to 1. 6 km. In regard to its age, it belongs wholly to the Carboniferous and does not go beyond the Tournaisian and Visean stages. The last member of the stratigraphic section — the upper volcanogenic formation — consists primarily of quartz porphyries, dacites, porphyrites and a large mass of coarse fragmental tuff-lava rocks.

The formation is distributed principally throughout the Shetsk and Aktogay areas and especially in the Northern Balkhash, from which it has also been given the name Balkhash formation [2]. In the Shetsk area the Balkhash formation overlies a Middle Visean coal-bearing suite, with closely associated transitional volcanogenic-sedimentary beds. Its upper chronological boundary lies in the Permo-Carboniferous. The total thickness is almost 4 km [1]. In the Balkhash area the formation begins either with the Visean or with the Middle Carboniferous; here its thickness has been given by various authors as 2 to 3 km.

The facies enumerated above, which represent natural paragenetic assemblages of rocks, are preserved over enormous areas from the village of Atasuy to Lake Balkhash, and maintain the same order of succession in the vertical section as well. Moreover they are not stratigraphic units in the narrow sense of the word, inasmuch as their ages differ in the various different areas.

All these changes in the chronological position of the formations are not random, but follow a regular pattern. As may be seen from the data set forth above and from the attached map, from northwest to southeast the formations gradually occur at higher and higher stratigraphic levels. For example, the Urtyn-Dzhal' formation of the Riphean and Cambrian merges into the Cambrian and Ordovician, the Kayraktinskaya formation moves from the Ordovician and Silurian to the Siluro-Devonian, the Sarysu-Teniz formation from the Lower and Middle Devonian to the Carboniferous (Tournaisian), and the Uspenskaya formation from the Middle Devonian and Tournaisian to the Tournaisian-Visean.

II. THE TECTONIC STRUCTURE AND DEVELOPMENT

The Uspenskiy zone is not a single tectonic structure, but consists of two subzones: the Uspenskiy basin proper and the Zhaksy-Tagaly thrust zone (Figures 2 and 3). The contours of the basin are determined mainly by the occurrence of the Uspenskiy (sandstone-carbonate) formation, which fills the basin. The structure of the thrust zone also includes the rocks of the Kayraktinskaya (sand-shale) and the Sarysu-Teniz (lower volcanogenic) formations.

1. The Uspenskiy basin directly abuts on the Tekturmass uplift and runs along it from the area of Atasuy to the village of Komkor, a distance of 245 km. At Komkor the basin is cut through by the Aktass fault zone, trending northwestward, and beyond this begins another basin, which extends from southeast to northwest and is filled with the rocks of the Balkhash (upper volcanogenic) formation.

The northern boundary of the Uspenskiy basin is, for the most part, not exposed. Along it lies the well-known Kaldyrma granite belt, which form an east-west bridge between the southern flank of the Tekturmass uplift and the central part of the Uspenskiy basin. The granites cut through the Uspenskiy formation and the Middle Carboniferous extrusives [20], and are considered to be Middle and Late Hercynian in age. The southern boundary of the Uspenskiy basin passes through certain reverse faults and thrusts that are disposed in an imbricated arrangement on the margin of the Zhaksy-Tagaly zone.

The form and dimensions of the Uspenskiy basin change from place to place, and it rises and falls along its longitudinal profile. In its extreme eastern part, between the Sherubay-Nura and Baygozha Rivers, the basin has the shape of a synclinorium some 20 km broad. On its flanks are exposed acidic extrusives of the Lower and Middle Devonian; in its center are sand-carbonate deposits of the Famennian and Lower Tournaisian, sand-shale deposits with carbonized plant detritus of the Tournaisian-Visean and acidic Upper Paleozoic extrusives; here, in fact, lies the large granite massif. The total thickness of the sedimentary deposits is 1.2 to 1.6 km, and that of the Upper Paleozoic extrusives is 200 to 300 m.

In the eastern part of the basin the rocks of the Uspenskiy formation are compressed into tight folds which morphologically closely resemble the folds in the older strata. The folds vary in width from a few tens of meters to 1 - 1.5 km, and the angles of dip are often as high as 70 to 80°. The large folds are accompanied by longitudinal faults which give them a step-like structure. On the northern flank of the synclinorium the folds are overturned toward the north, and on the southern flank toward the south. In places, under the pressure of the Zhaksy-Tagaly thrust zone, the folds of the southern half have become reoriented and are inclined northward. The southern flank of the synclinorium in such places is broken and somewhat shortened along its trend. As one approaches the Aktass fault zone, the trend of the folds changes from northeastward to almost straight north-south.

The central part of the Uspenskiy basin, between the Sherubay-Nura River and the area of the Uspenskiy mine, is a one-sided ramp. Its width in the east is about 15 km, and in the west 7 to 5 km. The section through the Uspenskiy formation in this area lacks the Tournaisian-Visean sandstone-shale deposits; the deposits below, in comparison to the corresponding ones in the eastern part, decrease to 100-150 m in thickness. Particularly widespread, especially near the Uspenskiy mine, are the Devonian extrusives, while in individual areas, probably along faults, there are exposures of

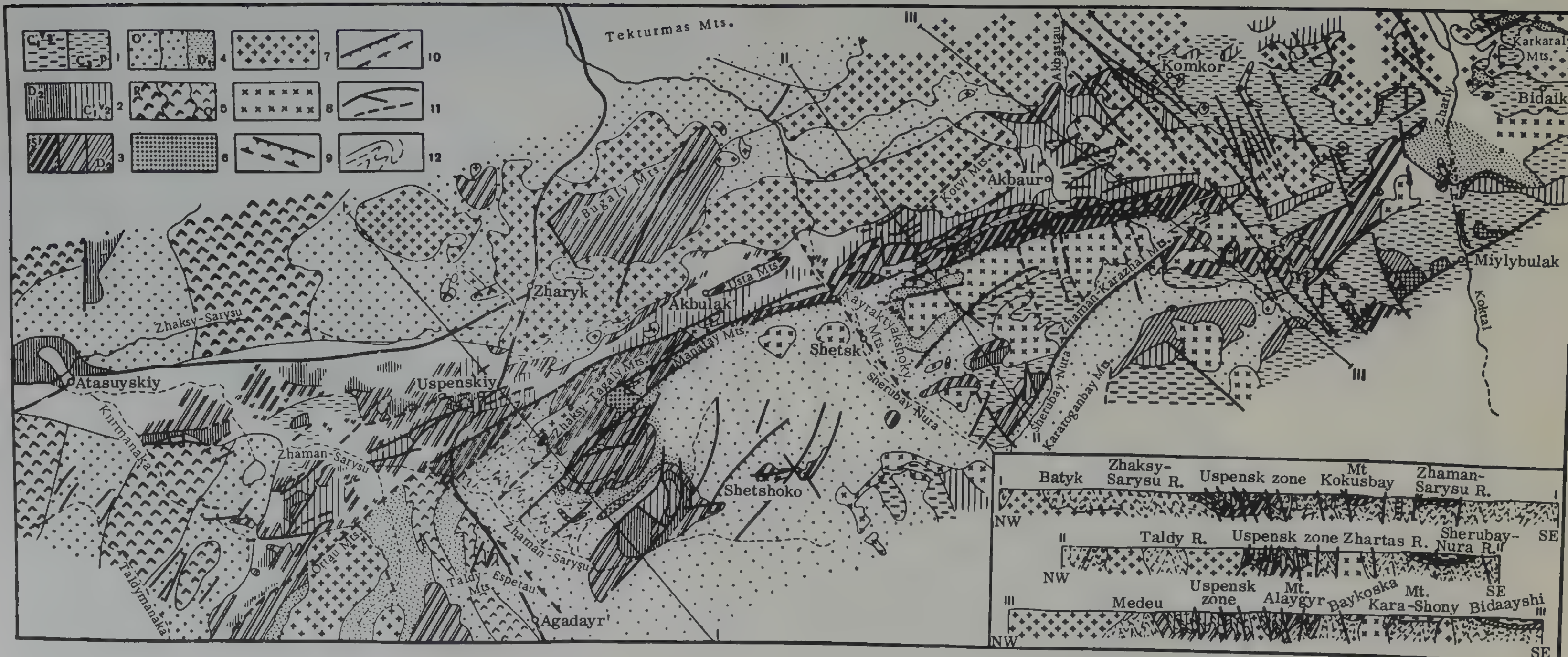

FIGURE 2. Diagram of the tectonic structure of the Uspensk zone and the adjacent regions of Central Kazakhstan (after A.I. Suvorov, 1959).

1 - Balkhash formation (Carboniferous-Permian); 2 - Uspensk formation (Upper Devonian, Lower Carboniferous); 3 - Sarysu-Teniz formation (Siluro-Devonian and Devonian); 4 - Kayrakta formation (Ordovician, Silurian, Lower Devonian); 5 - Urtyndzhal formation (Tiphean, Cambrian, Ordovician); 6 - hypabyssal intrusives (granite porphyries, quartz porphyries, syenite porphyries, quartz monzonites); 7 - Hercynian and Late Hercynian intrusives (biotite granites, alaskites, leucocratic granites); 8 - Early Hercynian intrusives (diorites, granodiorites, adamellites, quartz monzonites, granites); 9 - combined normal and strike-slip faults; 10 - combined reverse and strike-slip faults; 11 - other faults; 12 - outlines of folds and contacts between formations.

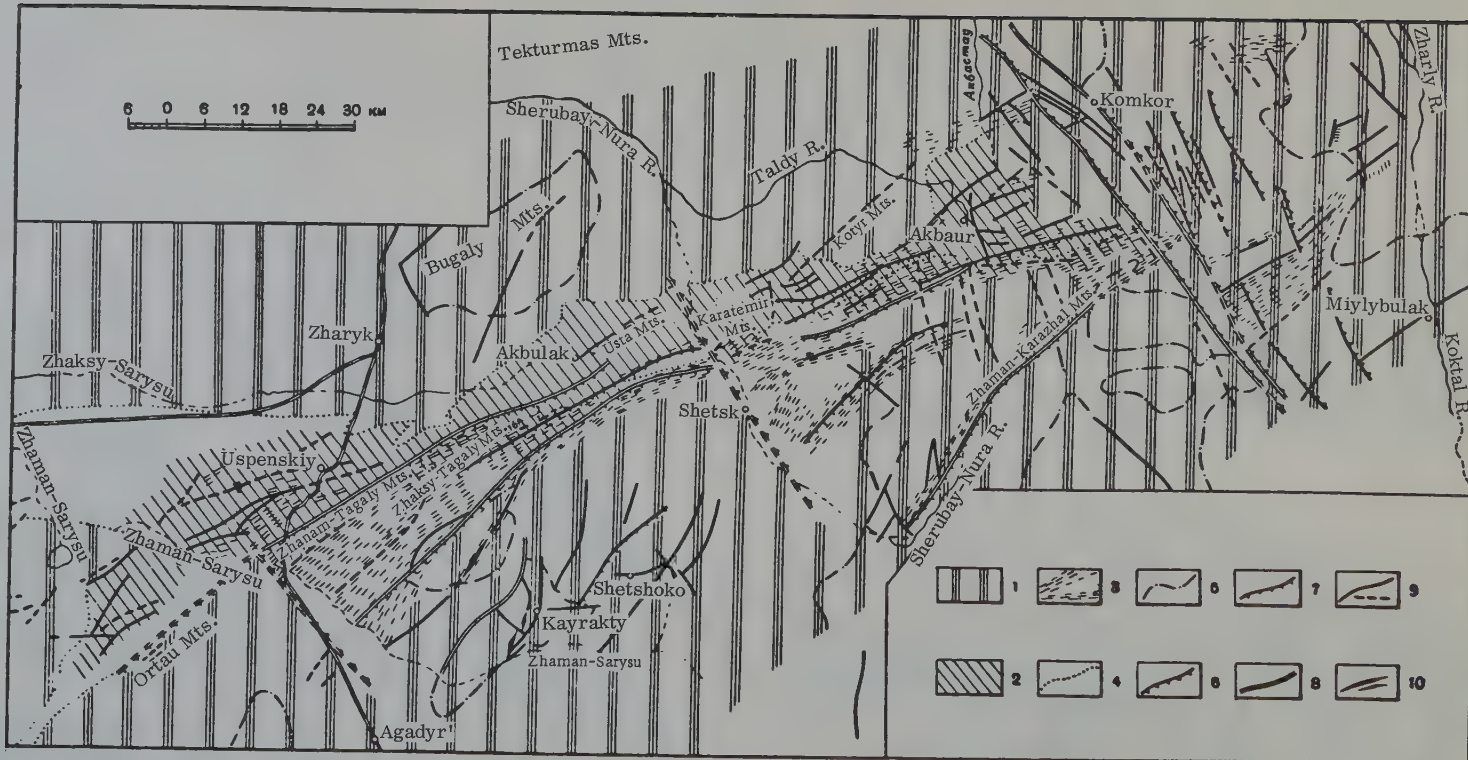

FIGURE 3. Sketch map showing the block-folded tectonics of the Uspensk zone and the adjacent regions of Central Kazakhstan.

1 - zones with block-folded structure; 2 - zone of compressive folding near a fault; 3 - zones of intensive cleavage; 4 - contours of the Uspensk basin proper, based on the $D_3 - S_1$ and in places S_2 deposits; 5 - contours of superimposed Middle and Upper Paleozoic basins; 6 - combined normal and strike-slip faults; 7 - combined reverse and strike-slip faults; 8 - faults contemporaneous with sedimentary deposition; 9 - faults occurring after sedimentation; 10 - axes of flexural folds.

Siluro-Devonian porphyroids and sand-shale deposits.

The folds in the sedimentary rocks are analogous to the folds in the eastern part of the basin, but for the most part they are inclined northward. Some of the reverse faults accompanying the folds emerge from the basin into the Zhaksy-Tagaly thrust zone, so that the southern border of the basin has a curious configuration. The structural forms of the eastern and central parts of the basin do not directly follow or continue each other. Between them there is supposed to be a northwestward trending fault, which runs along the valley of the Sherubay-Nura River.

West of the Uspenskiy mine the Uspenskiy basin broadens to 20 km and then to 40 km, and divides into two branches, one of which runs southwestward, following the trend of the older strata, while the other runs westward and overlies the older structures at almost a right angle to their trend.

The section through the western part of the Uspenskiy basin, besides the sand-carbonate deposits of the Tournaisian and Famennian, contains extensive red terrigenous deposits of the Frasnian stage and Devonian extrusives; G. I. Bedrov has, in addition, distinguished Carboniferous volcanic rocks. The total thickness of the Uspenskiy formation here, in comparison to the central part of the basin, increases to 1.5-2 km in some places.

Both branches of the basin are, in form, one-sided ramps, bounded on the south and southeast by reverse faults and steep thrusts. The folds that fill them are compressed and inclined northward. The interior faults are represented by comparatively small northeastward trending reverse faults, which are disposed at acute angles to each other in plan. Along these faults (as in the other parts of the Uspenskiy basin) the folds and the blocks of the Uspenskiy formation as it were ride up upon the older Tekturmass uplift.

2. The Zhaksy-Tagaly thrust zone has been traced by the present writer along the highly compressed and crumpled rocks of the Kayraktinskaya and Sarysu-Teniz formations from the Zhaman Sarysu River to the Aktass fault zone. It also continues farther northeastward for a certain distance, but is less distinct and is soon lost. The related compressed formations in the northeastern flank of the Aktass fault zone, which are of somewhat younger age, appear 15 km farther to the southeast.

The thrust zone is from 6 to 16 km broad, and consists of several long (up to 70 km) plates disposed en echelon on the map, each of them bounded on the northwest by a fault. There is no single fault that extends throughout the entire thrust zone. The faults take the form of reverse faults and steep thrusts, with a northwestward direction of movement of the rocks; they are readily distinguished from their localities, cut through various parts of the structure and are frequently accompanied by quartz veins and zones of quartz formation. Along these faults extends the narrow belt of volcanogenic rocks of the Sarysu-Teniz formation, with which it is genetically related.

Within this zone, the Kayraktinskaya and Sarysu-Teniz formations are generally folded conformably and oriented parallel to the reverse faults. The first of these contains carinate folds of several orders and small-scale plications. The folds in the porphyroids of the Sarysu-Teniz formation are larger, sometimes up to 3 km in width, but are also highly compressed. The flanks of most folds dip steeply, at angles of 60 to 70° or more. The axial planes of the folds are somewhat inclined toward the north and northwest.

The plates of rocks in some cases consist of several folds; in others they form monoclines or half-folds torn apart by the reverse faults. Within the plates there are also small faults in the form of short thrust or upthrust displacements, zones of fracturing and silicification. The plates also contain small intrusions of granites, granodiorites and diorites elongated northeastward. In comparison to the Kaldyrma granite zone on the northern edge of the Uspenskiy basin, these intrusives are more intermediate in composition. T. V. Perekalina and Ye. V. Petrova, who have studied some of the intrusives of the Zhaksy-Tagaly zone, consider them to be Early Hercynian.

Each of these plates of the Zhaksy-Tagaly zone has a somewhat different trend from the Uspenskiy basin and departs southwestward from it at a sharp and constantly increasing angle, also bending into an arc-like shape (see Figure 2). Thus the movement of the plates along the reverse faults was directed not exactly toward the Uspenskiy basin, but somewhat to the northwest of it at an acute angle. In some outcrops in the area of the compressed flanks of the northernmost reverse faults, composed of the limestones of the Uspenskiy formation, in the Akmay area, it has been possible to observe small horizontal folds whose orientation in plan testifies that the upthrust movements along northeastward trending lines were also accompanied by strike-slip movements directed westward along the Uspenskiy basin. Larger horizontal bends have also been noticed in the porphyroid rocks of the Zhaksy-Tagaly zone. The magnitude of the strike-slip displacement along one of these faults in the Zhaksu-Tagaly mountains, which separate this zone from the Zhasterek trough, has been determined as 1.5 km (with a vertical stratigraphic amplitude of no more than 1 km); it is supposed that there are

also other strike-slips here with amplitudes up to 3-3.5 km.

The rocks of the folded structures within the plates of the Zhaksy-Tagaly zone are intensively metamorphosed. The sand-shale deposits of the Kayraktinskaya formation have been turned into quartz-chlorite, quartz-sericite or quartz schists and sandstones. The volcanic rocks of the Sarysu-Teniz formation have undergone cleavage and are changed to porphyroids. These metamorphic alterations embrace the entire zone, but with varying degrees of metamorphism in different areas forming zones that are oriented parallel to the faults.

The maximal degree of dynamic metamorphism has been observed directly next to the displaced blocks of the upthrusts, especially those which border the extreme northwestern plates. The volcanic rocks here have completely lost their original structure and composition and in places cannot be distinguished from the quartz-chlorite or quartz-sericite schists of the Kayraktinskaya formation. Farther away from the upthrusts they are changes in appearance to porphyroids, and still father away simply show cleavage with secondary minerals formed only in the cleavage planes and, finally, at the farthest distance from the movements the dynamic action is reflected in the volcanic rocks only as very frequent jointing.

The longitudinal profiles of a number of plates of the Zhaksy-Tagaly zone are characterized by small uplifts and depressions. In addition the plates are sunk down toward the Uspenskiy basin, in the direction of which the folds sink downward and are covered by the younger deposits of the Uspenskiy formation, and in places also by Carboniferous volcanic formations.

Some of these depressions in the longitudinal profile in various plates move toward each other and mark transverse depressions of northwestward trend. Such, for example, is the depression in the Zhaksy-Tagaly Mountains, filled with the rocks of the Sarysu-Teniz formation. This depression or basin also encompasses the Zhasterek and Kayraktinskaya troughs and projects far to the southeast beyond the zone; its length is 35 km and its width 20 km. This category also includes the depression east of the Aktass fault zone, containing rocks of the Balkhash formation.

The transverse basins are associated with zones of transverse faults which cut both the Zhaksy-Tagaly zone and the Uspenskiy basin and extend into the combined Ortaushet-Shoka and Tekturmass uplifts. Such are the already described faults of the Aktass zone and the valley of the Sherubay-Nura River, as well as the fault along the valley of the Zhaman-Sarysu River and other smaller disjunctive dislocations.

Many of the northwestward trending faults have their northeastern flanks lowered and their southwestern flanks uplifted. At the same time the latter are displaced northwestward for a distance of several kilometers in some cases, and up to 10-15 km along the Aktass fault zone. These are apparently combined normal strike-slip faults. The normal and strike-slip nature of these faults was suggested as long as thirty years ago by M. P. Rusakov [21].

3. These deformations of the Uspenskiy basin and the Zhaksy-Tagaly thrust zone in no way resemble the deformations of the Tekturmass and Ortaushet-Shokinskiy uplifts. The Devonian and Carboniferous extrusives and sedimentary rocks preserved within these uplifts, which are contemporaneous with the Sarusy-Teniz and Uspenskiy formations of the Uspenskiy zone, in general have comparatively small, no more than 30 to 40°, angles of dip and form typical superimposed troughs, which are sharply set off from the tighter and more compressed structures of the basement complex by a surface of erosion and unconformity.

In the Tekturmass uplifts the superimposed troughs are the Sulu syncline and the Bugaly trough [3]; in the Ortaushet-Shokinskiy uplift they are the Zhasterekskaya, Kayraktinskaya, Chiyozekskaya and other troughs, which have been described at various times by the geologists of the Geological (GIN) and All-Union Geological (VSEGEI) Institutes. The imbricate and arc-like disposition of the faults in the Otraushet-Shokinskiy uplift, in comparison to the Uspenskiy zone, also causes the troughs to have corresponding orientations and dispositions.

The largest faults existed at the time of accumulation of the strata that fill the basins. For example, three depressions of the Zhaksy-Tagaly Mountains, which compose a basin trending northwestward and are separated from each other by northeastward trending faults, differ considerably in the composition of the strata of the same age. In the northernmost depression (a part of the Zhaksy-Tagaly zone) there is a stratum of Siluro-Devonian porphyroids that wedge out rapidly along the marginal fault of the Zhasterek trough, in which the Kayraktinskaya formation is transgressively overlain by younger extrusives of the Devonian. The upper part of the latter, which are of Frasnian age and are represented by albitophyric lavas, are replaced farther south in the Kayraktinskaya trough (after the next fault) by Frasnian sandstones in which the albitophyres are present only as a few small lenses. Here, too, appear carbonate-sandstone deposits of the Famennian and Tournaisian, of which the Kayraktinskaya trough is also mainly composed. In the troughs farther east, outpourings of the effusive rocks in the lower strata of the Balkhash formation are associated with certain of the faults that trend both northeast and northwest.

The area in which the superimposed troughs and the uplifts between them occur, in contrast to the Uspensk folded zone, is regarded as a zone of block-folded structure.

4. The formation of the Uspenskiy zone and the structures of the regions adjoining it took place against the background of a chronological rising displacement of the principal formational complexes from northwest to southeast. Their chronological migration began as early as the Riphean and continued throughout the entire Paleozoic. This can most readily be seen in the deposits which contain abundant fauna.

A complete geological interpretation of this extraordinarily interesting phenomenon is scarcely possible, so that at the present time it is possible to speak only of certain of its aspects.

As may be seen from the diagram showing the interrelationship of the sedimentary and volcanogenic formations (Figure 1), certain of these sometimes increase and sometimes decrease in thickness. It is suggested that this changing thickness is due to the formation of areas of subsidence and uplift. For example, the time during which the sedimentary rocks of the Kayraktinskaya formation were deposited saw the formation of three zones of subsidence (the Nura, the Shetsk and the Eastern Balkhash) and two zones of uplift (the Taldy-Espetau and the Northern Balkhash zones).

Since the formations become younger from northwest to southeast, the maximal and minimal thicknesses within each of them will, as one moves southeastward, occur at ever higher stratigraphic levels. Consequently in this direction the basins and uplifts were formed later, and the southeasternmost of them are the youngest.

The same picture appears in examining the paleogeographic data. On the facies diagrams drawn by A. A. Bogdanov [5], for example, the southeastern boundary of the dry land in Northern Kazakhstan was displaced 100 km southeastward during Ordovician and Gotlandian times, and on M. S. Bykova's diagrams [6], drawn for the Frasnian, Famennian, Lower Visean and Namurian stages, the Kyzyldzhar-Zharyk-Bayanaul area of erosion was displaced 80 km in this direction.

A study of the chronological migration of the formations leads to another conclusion as well: that the volcanic activity, and consequently also the faults with which the vulcanism was associated, also migrated in time — that is, that the fractures which served as channels for the effusive material were closed successively from northwest to southeast and were accompanied by ever younger tectonic dislocations. In certain of the largest fault zones the volcanic activity began earlier and continued for a longer time, as in the Zhaksy-Tagaly zone.

Finally, parallel to all this there was also a renewal of intrusive activity. For example, the granites of the Saryau-Teniz uplift are assigned by R. N. Sobolev and P. F. Yemel'yanenko [22] to the end of the Early and beginning of the Middle Devonian (the first complex) and to the end of the Middle and beginning of the Late Devonian (second complex). In the Uspenskiy zone, as was mentioned above, the granites are of Early Hercynian (after the Early Carboniferous) and Middle to Late Hercynian (after the Middle Carboniferous) age. Finally, in the Northeastern Balkhash region, according to V. G. Chuykova [26], appear the youngest granites, some of which cut across Upper Carboniferous deposits and others across the Lower Permian.

As a tectonic unit the Uspenskiy zone existed, probably, as early as the Silurian, bordering the earlier Tekturmass uplift as a synclinal basin of subsidence; its southwestern part, where sandstone and shale rocks of this age are exposed, is up to the present introduced, as it were, into the structures of the Urtyn-Dzhal formation, which follows the same directional trends. The zone had the form of a one-sided graben, bounded on the southeast by a series of faults arranged en echelon. Acidic extrusives, which formed a narrow belt along the southeastern margin of this graben, began to be poured out along these faults from the end of the Silurian. This effusive activity continued in the Devonian as well, embracing the interior parts of the Uspensk zone, especially in the west, where at that time a broad branch of this zone arose, superimposed at almost a right angle upon the meridionally trending structures of the eastern part of the Alasuy anticlinorium.

With the Late Devonian the faults along which the effusives had been pouring out were closed, and the Uspenskiy zone began to be an area of accumulation of predominantly sedimentary deposits. In the Late Devonian and the Carboniferous it underwent a small amount of subsidence and was separated from the other basins to the southeast (the Kayraktinskaya and Chiyozekskaya) by uplifted ridges, as indicated by the differences in the structures of their sections [1, 17]. The width of this basin of subsidence reached 40 to 50 km, and exactly this width is obtained if one conceptually restores to their horizontal position the intensively folded beds of the Uspensk formation. Volcanic activity was locally manifested in isolated places during the Late Paleozoic.

The folding within the Uspenskiy zone occurred most intensively during the Carboniferous and Permian, and was entirely associated with movements along faults. Most likely the folding began earlier, from the time the faults were closed as volcanic canals, when compressive forces began to be predominant.

The movements along the faults were oriented from southeast to northwest toward the Tekturmass uplift, in the direction opposite to that of the migration of the basins and uplifts. They took place along faults trending both northeastward (reverse-strike-slip faults and thrusts) and northwestward (combined normal and strike-slip faults), which were combined into a dynamic unity. Each southwestern area along the combined normal and strike-slip faults was elevated and displaced northwestward; along the reverse strike-slip faults and thrusts these areas were torn from their basement and also moved in the same northwestward direction. As determined by mapping, the magnitude of these displacements reached 10 to 15 km, but the maximal amplitude was considerably greater, if one recalls that in places the Uspenskiy basin was almost two-thirds covered (or pressed together) by the Zhaksy-Tagaly thrust zone.

Inasmuch as the movements occurred at an angle to the axis of the Tekturmass uplift, which acted as a fulcrum, certain of the block, folded and plate structures along the northeastward trending faults (reverse strike-slip faults) were somewhat displaced to the southwest. As a result, upon the Uspenskiy zone and the adjoining regions to the south, there was superimposed a system of "horse-tail" structures of the same nature as in other strike-slip-thrust and strike-slip zones. In this system the reverse strike-slip faults represent great shear fractures, which were formed under the conditions of predominant compression and were accompanied by parrallel elongated zones of intensive folding and dynamic metamorphism. The normal strike-slip faults are large tension fractures arising under the conditions of predominant tension, and serve as the boundaries of gently sloping superimposed troughs which are not accompanied by parallel zones of intensive folding.

III. SOME ANALOGUES OF THE USPENSKIY ZONE

The Uspenskiy zone, as an area of combined strike-slip and thrust structures, has many analogues, of which one, in particular, is the Cishissar zone in Tadzhikistan. This zone extends for 200 to 250 km and separates the structures of the Hissar Range from the Tadzhik depression; it was formed, in fact, as a result of the interaction of the latter two zones.

According to a number of authors, the Hissar Range is an upwarped zone trending approximately parallel to the equator, composed of Paleozoic granites and extrusive-sedimentary rocks, within which, primarily in the form of superimposed troughs, are preserved thin (up to 1.5 km thick) terrigenous and carbonate deposits of the Jurassic, Cretaceous, Paleogene and Neogene. The deformations here are exclusively of a block and block-folded nature and are due to the extensive occurrence of faults (reverse, normal and thrust) running east-west and northwestward.

In the Tadzhik depression the Paleozoic deposits are sunk to a considerable depth, and the Mesozoic and Cenozoic section even in its northern part is two or three times greater in thickness. Along the Hissar Range there is a basin filled with Cenozoic molasse deposits up to 5 or 6 km in thickness. Farther south, in the Mesozoic and Cenozoic deposits, one sees a large number of linear folds, monoclines and plates elongated and extending generally from southwest to northeast for tens of kilometers, each of them some 5 to 7 km wide. Figure 4 shows that all the structures (as in the Uspenskiy zone) form arc-shaped patters convex northwestward, projecting from the structures of the Hissar Range and the Cishissar basin at an acute and constantly increasing angle.

In the eastern part of the depression, as is known from many papers, the folds are overturned and displaced farther to the northwest along a system of longitudinal northeastward trending thrusts, the largest of which are the thrust of the Ilyak River valley and the Vakhshskiy thrusts mentioned by I. Ye. Gubin [11 et al.]. The entire space between these two dislocations in the vertical section has a structure in the form of plates sliding over the structures of the Hissar Range and the Cishissar basin. East of the city of Fayzabad the basin is entirely covered by these thrust plates, from which it is determined that the smallest magnitude of the thrust displacement was about 20 km. A small role was probably also played here by the northwestward trending faults along which displacement took place, such as the Tutkaul'skiy underground fault (Figure 4).

The ubiquitous eastward bending of the structures of the Tadzhik depression as one approach as the foothills of the Hissar Range is accompanied by east-west faults. M. S. Shvetsov [27] called attention to this fact a long time ago. Later S. A. Zakharov [12, 15] showed that these faults are strike-slips with displacements of their southern blocks westward for distances of from several hundreds of meters to a few tens of kilometers.

Thus in the Cishissar zone were of the same strike-slip and thrust character as in the Uspenskiy zone, and determined not only the similar tectonic structure of both zones but also the morphological relationship of the particular structural forms within them. Both zones, on the other hand, are combinations of large obliquely oriented structures with basins at the ends of the latter. The fact that they belong to different structural stages indicates that such zones may occur in geologic formations of any age. When a sufficient number of exsamples have been studied, it will ultimately be possible to

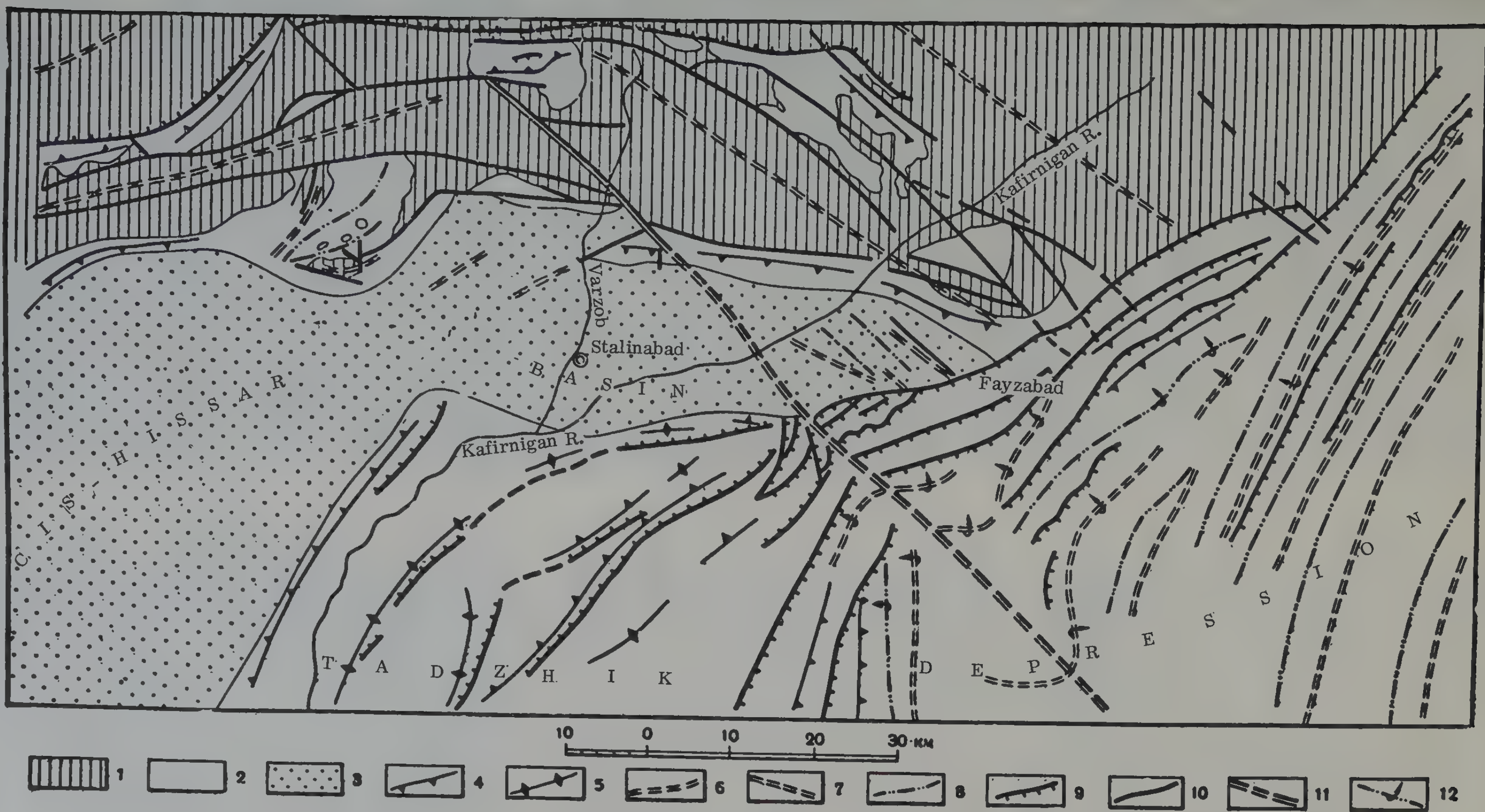

FIGURE 4. Sketch map showing the structure of the Hissar Range and the adjacent parts of the Tadzhik depression (after A.I. Suvorov).

1 - Paleozoic; 2 - Mesozoic and Cenozoic; 3 - areas of greatest thickness of Cenozoic molasse deposits (up to 5-6 km); 4 - large monoclines; 5 - box-shaped anticlines; 6 - flesural anticlines and block-uplifts; 7 - compressed anticlines (comb-shaped and others); 8 - axes of synclines; 9 - thrust faults; 10 - reverse and normal faults and other faults in the basement complex; 11 - Tutkaul' buried fault; 12 - direction of overturning of the structures.

establish the entire range of their chronological changes.

It is interesting that the series of the features of strike-slip and strike-slip-thrust zones may be observed in zones of different scale and order of manifestation. In this sense it is worth while to compare the Uspenskiy zone with the larger zones of the Alpine strike-slip in New Zealand and the Talass-Fergana strike-slip in Central Asia.

The Alpine strike-slip zone occurs on South Island and extends northeastward at 50° for 800 km. According o H. W. Wellman [33, 34], the structure of this zone includes Precambrian, Paleozoic and Mesozoic and Cenozoic rocks, which form several structural stages or levels. Each successively younger structural-facies zone arose on the Pacific Ocean (that is, eastern) side of the older zone.

The Precambrian and Early Paleozoic rocks of South Island consist of gneisses, marbles, quartzites, graywackes and shales, which are located primarily in the northwestern block of the strike-slip fault. In the Carboniferous and Permian flows and sills of basalts, andesites and spilites predominate; these rocks occur in the northeastern and southwestern parts of the Island and, together with the Cambrian rocks, have been moved apart by the strike-slip to a distance of up to 480 km. The Upper Paleozoic, Triassic and Jurassic is represented by graywackes, conglomerates and limestones; of the same age (Carboniferous to Middle Triassic) are a thick series of schists. All these rocks are located mainly southeast of the Alpine strike-slip. Directly along the strike-slip fault are gneisses formed from pre-Cretaceous granites.

The metamorphic and sedimentary formations of the Upper Paleozoic and Lower Mesozoic in the southeastern block of the strike-slip fault are crumpled into folds which project southwestward from the fault plane at an acute and ever increasing angle, thereafter forming an arc with its convex side facing west. According to the detailed investigations made by Lillie, Gunn and Robinson [32], the folds in the metamorphic complex are steep and isoclinal with their axial planes dipping southeastward in the strike-slip zone and northwestward farther away from it. As one approaches the strike-slip there is a sharp increase in the degree of metamorphism of the rocks and garnet-biotite schists appear, which contain micro-folds, bounding structures cleavage in two directions, etc., while to the east there are rocks of the chlorite subzone in which there is only cleavage parallel to the bedding planes. Closer to the fault there are also numerous longitudinal and transverse faults that break the rock into blocks which rise like steps from east to west. In the northeastern part of South Island H. Wellman mentions a group of intersecting faults trending east-northeast, the southern blocks of which are regularly displaced westward; the amplitude of the displacement along some faults is as much as 20 km.

Evidently the southeastern block of the Alpine strike-slip, composed of younger formations, was at the same time the more mobile. It was broken up and bent throughout the whole of South Island, displacing mainly westward. Characteristically the movement of the rocks of this block, as in Kazakhstan, was in the direction opposite to that of the chronological migration of the structural-facies complexes.

The Talass-Fergana strike-slip, according to V. N. Ognev [18] and other authors, separates the ancient Caledonian mass from the area of Hercynian structures with its thick Middle Paleozoic strata. Farther southeast, the strike-slip fault cuts into the latter, interrupting and displacing its interior zones.

The area to the northeast of the fault was for a long time the site of vertical, primarily upward movements, along with the long-lasting formation within the same territory of large basins and uplifts of block-folded structure. In the area of the opposite side of the fault, on the other hand, downward movements predominated, and the many basins and uplifts had a brief period of development and migrated in time from one area to the other [25]. This fault block was the more mobile, and was displaced along the strike-slip fault in the northwestward direction. The magnitude of this displacement has been determined by V. N. Ognev as 150 km on the average, whereas according to L. B. Vongaz [8] the amplitude of the strike-slip decreases from 180 km in the northwest to 90 km in the southeast.

The mobile southwestern block was the more intensively folded. The Paleozoic deposits here contain a series of folds and faults which project out from the main fault at acute angles toward the northwest and then bend into an arc. In the direct vicinity of the fault the folds are isoclinally compressed, and the sandy-clay and carbonate rocks, according to V. N. Ognev, are mylonitized and metamorphosed to mica schists and marbles. The Mesozoic deposits contain superimposed folds and steep autonomous folds. In the most highly compressed folds at the fault the clay rocks are transformed into schists and the coals into anthracites. Farther from the fault, in Eastern Fregana, there are extensive manifestations of tear deformations produced in the thrusting of the Mesozoic and Cenozoic mantle over the Paleozoic basement (for the most part) northward and northeastward [24]. The movement of the younger formations over the older that is associated with the strike-slip here, as in the other zones, took place in the opposite direction to that of the successive arising uplifts and basins, which in Eastern

Fergana migrated with time from northeast to southwest and west [14, 19, 24].

These data cited for comparison show that strike-slip and strike-slip-thrust zones are characterized by the following common features. Each zone divides areas of different structures and developments, of which one is apparently younger and more sunk down and is the more mobile, while the other is older, less mobile and massive. The first is allochthonous and the second authchthonous.

In the mobile allochthonous part a series of very compressed folds and faults develops; these are disposed in plan relative to the edge of the authchthonous sheet as horizontal plates that are displaced and curved arc-wise always to the left. Among the faults are very characteristically conformable thrusts and reverse faults that are closely associated with the strike-slip, along which the individual plates slip over each other or are steeply displaced upon each other and upon the authochthone (or in its direction). There are frequent transverse normal strike-slip faults which cut through the plates and displace the parts of these plates also in the direction of the autochthone. The rocks of the front edge of the allochthone undergo a high degree of dynamic metamorphism, which is to be observed in almost all of the zones under consideration. The uplifts and basins in the area of the allochthone are characterized by a brief period of development and migrate in time from place to place. The mechanical movement of the mass of rocks along the strike-slip surface takes place in the direction opposite to that of the successive appearance of the uplifts and downwarps in time.

Evidently all the strike-slip and strike-slip-thrust zones enumerated in this article belong to the same genetic type of tectonic dislocation — the type of right strike-slips with simultaneous development of reverse and thrust displacements. The differences between them depend mainly on the age and on the order of their manifestation. An important factor also is the magnitude of the angle between the vector of mechanical displacement of the strata and the trend of the uplifts and basins arising in the zone. The smaller this angle, the more clearly manifested will be the strike-slip effect proper (New Zealand); the closer this angle is to 90°, the greater will be the role of the thrusts and reverse faults (Uspenskiy zone). All these criteria may form the basis of a future classification of the dislocations of this type.

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ON THE SUBDIVISIONS OF THE LOWER CAMBRIAN[1]

by

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Until recently, many investigators have used the system by which the Cambrian deposits are subdivided into stages on the Siberian platform as the basis for their world-wide correlation of the Lower Cambrian subsystem. In the type section, or so-called "eastern" section along the Lena River the Lower Cambrian, according to the changes in the olenellids and the accompanying Triangulaspis, Hebediscus, Neocobboldia and Pegetiellus protolenellids, is clearly divided into two stages: the Aldanian and the Lena stages [4, 7].

Nevertheless the extension of these subdivisions to a greater area encounters serious difficulties. In the Altay-Sayan folded region, for example, it has turned out that the entire Lower Cambrian is represented by the Lena stage [2], in spite of the fact that it is conformably underlain by strata with an assemblage of algae typical of the Riphean. In North America, on the other hand, the olenellids extend upward to the Middle Cambrian [14, 15] and here, consequently, the Lena stage is entirely absent. At the same time it has been shown that the Middle Cambrian, both in Siberia and in North America, in many sections merges with the Lower Cambrian in a gradual transition and begins with beds of the same age [4, 6, 10, 14, 15].

The data obtained recently on the stratigraphy of the Lower Cambrian in the northern region of the Siberian platform and the elaboration of the biostratigraphy of the deposits of this age in the Sayan-Altay folded region has drastically curtailed the territory in which the subdivision of the Lower Cambrian into the Aldanian and Lena stages can be considered as reliable, and at the same time have suggested other approaches to the subdivision and correlation of the Lower Cambrian deposits of Siberia.

On the basis of the paleontological data, the Lower Cambrian of the Altay-Sayan folded region, which is subdivided into five levels [2], can be correlated in great detail with the Lena stage of the "western" type section on the Siberian platform (the western sections on the Tolba and Olekma Rivers and the Irkutsk amphitheater); (see the accompanying stratigraphic table).

In correlating the Lower Cambrian section of the Sayan-Altay folded region with the "eastern" (the type) section along the Lena River, it appears that according to the archaeocyathid assemblages (data from I. T. Zhuravleva and A. Uy. Rozanova) the Bazaikhskiy and Kameshkovian[2] levels of the former correlate with the Kenyadinskiy and Atdabanskiy levels in the Aldanian stage of the latter. The occurrence of trilobites confirms this correlation. Thus the genus Tumulina, which is characteristic of the Kameshkovian level, has been found in a number of sections in the multicolored formation of the Aldanian stage on the Yudoma River. Along the Botoma River the beds that grade into the Lena stage have been found to contain such typically Sanashtykgolian forms as Botomella and Rondocephalus. On the other hand, olenellids (Fallotaspidella) are known in the Sayan-Altay folded region from the beds with Kameshkovian fossils. The other forms of the Aldanian stage — Hebediscus, Triangulaspis and Pagetiellus — here do not occur above the top of the Sanashtykgolian stage. These data compel some doubt regarding the correctness of the correlation between the "western" and "eastern" type sections, the more so because the parallels drawn between their lower strata have no paleontological basis.

It is apparent that some lower part of the paleontologically identified Lower Cambrian of the Sayan-Altay folded region, as well as the

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[1]O podrazdelenii nizhnego kembriya, (pp. 83-87).

[2]Earlier [2] the Kameshkovian level had by the present writers below the Bazaikhskiy level. In recent years a characteristic assemblage of Kameshkovian fossils has been discovered in continuous sections between the Bazaikhskiy and the Sanashtykgol'skiy levels; this discovery fixes the stratigraphic position of the Kameshkovian strata.

corresponding strata of the "western" type of section of the Siberian platform, are correlated with the Aldanian stage of the "eastern" type section. Thus in the Lower Cambrian sections in the Sayan-Altay folded region and in the western sections on the platform, the boundary between the Aldanian and Lena stages is not marked by any considerable change in the fossil community.

On the other hand, throughout this entire territory there is a sharp shift in the development of the archaeocyathid-trilobite fauna in the Lower Cambrian rocks. In the Sayan-Altay region this runs along the boundary between the Sanashtykoglian and Solontsovian levels [3]. In the platform sections this shift in the fauna corresponds to the boundary between the Angara and the Botomayskian substages,[3] along the top of the Olekma level (which is of the same age as the Sanashtykgolian stratigraphic level). A similar picture is to be observed in the northern part of the Siberian platform.

In the sections through the Cambrian rocks of the Anabar shield and the Olekma uplift, the geologists of the All-Union Scientific Research Institute for the Geology of the Arctic distinguish the Aldanian and the Lena stages and believe that these are represented in their full stratigraphic extent [9]. In most of the sections, however, the entire group of strata that belong to the Lena stage has a total thickness on the order of 5 to 10 m, whereas the part of the section that contains the fossils of the multicolored formation and is assigned to the Aldanian stage reaches several hundreds of meters of thickness. The bituminous shales of the Lena stage have been found to contain Lermontovia grandis and Granularia obrutchevi, which occur in the upper layers of the Lena stage on the Lena River. In addition, the bituminous shales also contain Lermontovia dzevanovskii and Bergeroniellus asiaticus; on the basis of these fossils it is supposed that the section contains the lowermost Sinskian level of the Lena stage. In the Lena River sections, however, the occurrence of Bergeroniellus asiaticus rises to the top of the Lower Cambrian. Lermontovia dzevanovskii moreover so closely resembles L. grandis that it is often indistinguishable. In a number of sections in the Anabar anteclise, both species occur together. For these reasons the presence here of the lower strata of the Lena stage is not firmly established, and it is not impossible that the whole Lower Cambrian part of the bituminiferous formation and its analogues are to be correlated only with the Yelanskian and, perhaps, the Ketemenskian levels, and that the upper part of the underlying deposits, which are parallel to the multicolored formation, are correlated with the lower part of the Lena stage on the Lena River. This interpretation is directly supported by the fact that in the upper part of the Aldanian stage as distinguished here, together with the fossil forms characteristic of the multicolored formation — Triangulaspis, Hebediscus, Neocobboldia and the Olenelliade — there is the guide fossil of the Olekma level, Jacutus quadriceps, as well as Calodiscus, which as a rule occurs in the upper half of the Lower Cambrian. In the Igarka region along the Sukharikha River (a right tributary of the Yenisey River), according to a verbal communication from V. I. Dragunov, together with Jacutus, Botomella and Calodiscus (forms characteristic of the Olekma and Sanashtykgolian levels) there are the Olenellidae, Neocobboldia and Pagetiellus, which belong to the Aldanian stage. Lermontovia is present higher up in the Severorechinskaya formation.

In the northern part of the Siberian platform, consequently, the only sharp change in the fossil assemblage corresponds to the top of the Olekma level — that is, it occurs much higher stratigraphically than in the type section. It coincides approximately with the main boundary in the Lower Cambrian sections of the Sayan-Altay folded region and to the boundary between the Botomayskian and Angarian substages on the Siberian platform.

Thus throughout almost all of Siberia, two major stratigraphic subdivisions are distintuished within the Lower Cambrian.

The situation is otherwise with the subdivisions identified according to the replacement of the ollenellids by the protolenids or the Aldanian and Lena stages. In the type section the change between these two fossil communities occurs at the base of the Sinskian level, where the multicolored formation is known to be replaced suddenly, higher up in the section, by the bituminous shales and limestones.

In the Anabar, Olenek uplift and Karaulakh sections and, apparently in the basin of the Yudoma River, the occurrence of the olenellids and the assemblage associated with them rises to the top of the Olekma level, where the multicolored rocks are replaced in a sharp transition by the bituminous limestones and shales [1].

In a number of sections of the "western" type on the Siberian platform the protolenids occur in considerable quantity only at the Olekma level (according to N. P. Suvorova, [8]). But in most of the areas that have sections of the "western" type, as well as throughout almost all of the Altay-Sayan folded region, the olenellids and protolenids are absent and it is almost impossible to determine where to draw the

[3]In the "western" type section this is the first important boundary; in the "eastern" section it is no less clear, but is overshadowed by the sharp change in the fossil fauna at the boundary between the Aldanian and Lena stages.

A CORRELATION OF THE LOWER CAMBRIAN SECTIONS

<table>
<tr><td colspan="3" rowspan="2"></td><td colspan="2">Altay–Sayan folded region</td><td colspan="4">Siberian platform</td></tr>
<tr><td>Eugeosyncline, after I.T. Zhuravleva, L.N. Repina and V.V. Khomentovskiy, 1959</td><td>Miogeosyncline, after V.V. Khomentovskiy, M.A. Semikhatova and L.N. Repina, 1960</td><td colspan="2">Aldan shield, Olekma River, after N.P. Suvorova, 1960</td><td colspan="2">Western margin, Sukharikha River, after V.I. Dragunova</td></tr>
<tr><td colspan="3">Middle Cambrian</td><td></td><td></td><td colspan="2"></td><td colspan="2"></td></tr>
<tr><td rowspan="8">Lower Cambrian</td><td rowspan="2">Upper subseries</td><td>Obruchevian level</td><td>Edelsteinaspis, Chondragraulos, Kooteniella, Erbia, Batenioides, Pumulina</td><td>Edelsteinaspis Chondragraulos Kooteniella Erbi</td><td>Yelanskian level</td><td></td><td colspan="2" rowspan="2">Severnaya River formation
Lermontovia
450 m.</td></tr>
<tr><td>Solontsovian level</td><td>Onchocephalus, Onchocephalina, Laminurus, Solontzella, Proerbia torgaschinica</td><td>Pseudoeteraspis Parapoliella Bathiuriscellus, Proerbia torgaschinica</td><td>Katemenskian level</td><td>Charskian formation
Nomanoia, Batynotus
200 m.</td></tr>
<tr><td rowspan="6">Lower subseries</td><td>Sanashtykgol'skian level</td><td>Rondocephalus, Poliellina, Poliellaspis, Botomella, Serrodiscus, Calodiscus, Redlichina, Erbiopsis, Jnouyina</td><td>Tungusella Kolbinella Bigotina Bulaiaspis sajanica, Jnouyina, Termieroides</td><td>Olekma level</td><td>Olekma formation
Jakutus, Solenopleurella, Bergeroniaspis Tungusella
100 m.</td><td rowspan="4">Sukharikha formation</td><td>Calodiscus, Jakutus, Botomella, Pag i llus, Neocobboldia, Olenellidae</td></tr>
<tr><td>Kameshkovian level</td><td>Sajanaspis, Palaeolenella, Hebediscus, Planaspis Fallotaspidella, Bulaiaspis taseevica</td><td>Bulaiaspis taseevica, B. vologdini, B. prima</td><td>Tolbachanian level</td><td>Tolbachan formation
Bulaiaspis
200 m.</td><td rowspan="3">750 m</td></tr>
<tr><td rowspan="2">Bazaikhskian level</td><td rowspan="2">Resimopsis Kijanella Lunolenus Elganellus Mitioides Paraerbia</td><td rowspan="4"></td><td>Sinskian level</td><td>El'gyan formation
Elganellus, Malykania
60 m.</td></tr>
<tr><td colspan="2">Yuyedeyskaya formation</td></tr>
<tr><td rowspan="2"></td><td rowspan="2"></td><td colspan="2">510 m.</td><td colspan="2" rowspan="2"></td></tr>
<tr><td colspan="2">Tolbinskaya formation</td></tr>
</table>

A CORRELATION OF THE LOWER CAMBRIAN SECTION (continued)

	Anabar shield (eastern slopes) after V.Ye.Savitskiy and K.K.Demidkov, 1959	Africa, Morocco, after P.Hupe, 1952	North America				
			Mexico, after C.Lockman, 1956			New York State, after C.Lockman, 1959	
Kounamian level	Anabaraspis, Poulsenia, Paramicmacca, Kooteniella, Granularia, Lermontovia Grandia, Lermontovia dzevanovskii, Bergeroniellus 5—7 m	VIII Protolenus, Lusatiops, Homatolenus, Pseudolenus, Collyrolenus, Myopsolenus, Kingaspis, Micmacca 60 m	Upper Olenellus subzone	Antagmus-Onchocephalus	Sierro Prieto formation Girvanella 110 m.		
		VII Protolenus, Lusatiops, Micmacca, Kingaspis,			Buelna formation Onchocephalus, Antagmus, Bonnia, Olenellus, Paedumias 40-100 m.		
Ukhumunian level	Hebediscus, Triangulaspis, Calodiscus, Jakutus, Pagetiellus, Olenellidae 11 m	Serrodiscus, Termierella, Gentilaspis 250 m		Providora formation Olenellus, Linguloides 250 m.		Elliptocephalla asaphoides	Antagmus, Atops, Elliptocephalla, Serrodiscus, Calodiscus, Bonnia, Fordaspis, Bonnaria, Kootenia, Labradoria
		VI Callavia, Holmia, Termierella, Saukianda, Resserops, Eops, Perrector, Longianda 200 m					
Yulegir-Yuryakh	Pagetiellus, Hebediscus, Paedumias, Hyolites 50 m	V Bondonella, Strenuella, Pruvostinoides, Weymouthia 150 m					
		IV Fallotaspis, Hebediscus, Pareops 130 m					
Chabur, Luchatka	100 m. Archeocyathids of the Kenyadinskiy level	III Fallotaspis, Pruvostina, Ouijjania, Despujolsia, Neoredlichia 200 m					
		II Fallotaspis, Choubertella, Pruvostinoides 50 m	Lower Olenellus subzone	Puerto Blanco formation			
	60 m	I Fallotaspis, Bigotinops, Pararedlichia 50 m		Wanneria, Olenellus, Obolella 320 m			

boundary between the Aldanian and the Lena stages.

The data presented above suggest that the replacement of the Aldanian by the Lena assemblage of fossils was due to local causes. In Siberia in particular this is to a great degree determined by the facies changes.

The range of the fluctuations in the level at which the olenellids are replaced by the protolenids — taking account of North America, where this change occurs as high as the bottom of the Middle Cambrian — is too great to allow the subdivisions of the type of the Aldanian and Lena stages to be used for correlating Lower Cambrian deposits over extensive territories [5, 11]. Despite the fact that in some places (the northern part of the Siberian platform, for instance) the level at which the olenellids give way to the protolenids coincides by chance with the boundary between these two subdivisions, the two have nothing to do with each other. To stress the independence of the two major stratigraphic units, distinguished on the basis of the changes in fossil assemblages which differ greatly in different places, the present writers will here, with some degree of tentativeness, call them subseries. The use of such a term for these subdivisions is justified by the fact that they can be traced in many Lower Cambrian sections throughout the world. In North America, for example, the subdivision of the Lower Cambrian generally accepted at the present time envisages an upper and a lower Olenellus subzone; but many authors [14, 15], near the top of the upper Olenellus subzone, distinguish an independent subdivision characterized by the occurrence of a multitude of new genera of trilobites that are not encountered below, as well as by the almost complete disappearance of the olenellids. The deposits of this part of the section is combined by C. Lochman [14] are grouped together into the packet of beds containing Antagmus — Onchocephalus. It is very interesting that the trilobite assemblage in these beds closely resembles that of the Solontsovian level in the Sayan-Altay region (the Ketemenskian platform), which also contains such widespread forms of the Antagmidae family as Onchocephalus, Onchocephalina, etc.; on the basis of these genera, it may be said that the North American and Altay-Sayan sections are of the same age. In North America one observes the same considerable change in the trilobite community at the same stratigraphic level. The same conclusions are suggested by a study of the Lower Cambrian sections in Morocco. Here the Lower Cambrian is subdivided by P. Hupé [12] into eight zones; here the olenellids occur quite high in the section, being found in the six lower zones, occurring rarely in the seventh and being completely absent from the uppermost, the eighth, zone of the Lower Cambrian rocks. On the other hand, here the protolenids, represented by six genera, are widespread. This part of the section evidently corresponds to the Yelanskian and Ketemenskian levels in Siberia and to the Antagmus — Onchocephalus beds in North America. The confirmation of this is the close similarity of the underlying complexes to those of the Sayan-Altay region. For instance, the sixth and seventh zones in Morocco contain such forms as Termierella, which is close to the Olekma form Termieroides, and Serrodiscus, which occurs in the Sanashtykgolian assemblage of the Sayan-Altay region, and so forth.

The trilobite assemblages from the lower four zones closely resemble the Kameshkovian and Bazaikhskian complexes of the Sayan-Altay region. More specifically, the genus Fallotaspis is close to Fallotaspidella; Ouijjania very strikingly recalls the genus Minusella, etc. It is interesting that representatives of the family Neoredlichiidae occur in this part of both sections. One must also take into account the assemblage of trilobites found in the Lower Cambrian deposits in the border area between the Atlantic and Pacific Provinces, near Hoosich in the state of New York [13]. C. Lockman has named this assemblage after the most typical representative of the olenellid family, Elliptocephalla asaphoides; it contains many forms common to the Sanashtykgolian complex in the Sayan-Altay region, including Serrodiscus, Calodiscus, Bonnia and others. Lochman has correlated the bottom of the Elliptocephalla asaphoides beds with boundary between the lower and upper Olenellus subzones. But the distribution of the forms in the Elliptocephalla asaphoides beds, as cited in the summary article by C. Lochman and J. Wilson [15], shows that they all occur below the lower boundary of the upper Olenellus subzone. Thus one has the right to say that the beds containing Elliptocephalla asaphoides, and consequently also the Sanashtykgolian level (the Olekma) of the Sayan-Altay region, are correlated with the middle part of the Upper Olenellus subzone in America.

Characteristically, the representatives of the olenellid family and the associated forms occur up to the very top of these strata, rarely occurring above, whereas new assemblages appear to take their place. This boundary marking the replacement of one assemblage of organic remains by another is observed in many sections of the Lower Cambrian throughout the world and is broken off in the Pacific and Atlantic Provinces.[4] Moreover, as has been seen, in various different countries one observes approximately the same succession of fossil assemblages, many of which contain similar forms through which they can be connected.

Thus the Lower Cambrian sections that have been considered in the Atlantic and Pacific

[4]The East Asian province is not considered here.

provinces are made up of the same two subseries as in Siberia.

Within the much larger lower subseries, it is already possible to trace subordinate but quite clearly connected smaller subdivisions. For example, in the upper part of the lower subseries, there is a subdivision corresponding to the Elliptocephalla asaphoides beds of America, the sixth and seventh zones in Morocco, and the Sanashtykgolian and Olekma levels of Siberia. Lower levels of this subseries are as yet still not universally correlated.

The lower parts of the sections through the Lower Cambrian are, as a rule, only to a slight degree characterized by fossil contents (they have not been considered above). These contain rare olenellids, pteropods and brachiopods. This part of the Lower Cambrian section may very tentatively be set apart as an independent subseries. In Siberia this includes the lowest strata of the multicolored formation, the upper part of the Tolbinskaya formation, and the Yudomskaya formation and its analogues, which contain pteropods isolated single, or often unconfirmed collections of olenellids. In America this part of the section includes the lower beds of the Lower Olenellus subzone (organic remains usually are found in the upper part of this subzone). In Africa it is represented by the Adudu Limestone beds, which contain sparse archaeocyathids and algae, and apparently also the first zone distinguished by P. Hupé, which contains ancient archaeocyathids. It will be seen from the above that the organic remains in these strata are so sparse as to provide no data for reliably distinguishing these beds from the Riphean group.

The subdivision of the Lower Cambrian into subseries requires a far stronger basis in data than it has at present; thus the stratigraphic scheme presented in this paper is to be taken only as a preliminary one.

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NEW DATA ON THE STRATIGRAPHY OF THE LOWER JURASSIC MARINE DEPOSITS ALONG THE VILYUY RIVER[1]

by

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Much attention has been devoted to the detailed study of the stratigraphy of the deposits of combustible gas and of oil shows in certain districts of Yakutia. These deposits, which fill the enormous areas of the Vilyuy syneclise and the Verkhoyansk marginal basin, include extensive Jurassic sediments. The areas in which these deposits occur, especially in the vicinity of the Paleozoic, are also of interest from the practical standpoint of prospecting for possible placer and bedrock deposits of diamonds. Stratigraphic studies also play a large role in geologic mapping and in the preparation of the State Geologic Maps.

A BRIEF HISTORY OF THE STUDY

Along the Vilyuy River the Jurassic marine deposits have been studied for a long time. In 1853 - 1854 R. K. Maak [11] first mentioned the presence of Jurassic marine deposits along the Vilyuy River. In 1861 a member of his expedition, Pavlovskiy, continued the investigations along the banks of the Vilyuy River and determined that the Jurassic deposits extend from the Vilyuy River eastward in the direction of the Kempendyaya River.

In 1917 the Vilyuy was visited by the famous geologist A. G. Rzhonsnitskiy [12], who worked out the first stratigraphic scheme for the Jurassic deposits, within which he distinguished the following: Liassic fresh-water deposits, marine deposits of the Lower Doggeran, and Upper Jurassic fresh-water deposits. In the Jurassic deposits this investigator discovered an ammonite, which was subsequently identified by A. P. Pavlov as the Upper Aalenian Harpoceras murchisonae Sov. A. G. Rzhonsnitskiy's stratigraphic scheme was widely used in its time.

A more precise account of the stratigraphic subdivision of the Jurassic deposits occurs in the joint work by G. A. Krymgol'ts, G. T. Petrova and V. F. Pshelintsev [10], who, on the basis of samples collected from Yakutia, noted the following subdivision for the basin of the Vilyuy River: Liddle Liassic, Upper Liassic and Upper Aalenian with Ludwigia murchisonae Sow. The chief defect of this scheme is that these subdivisions included strata of different stratigraphic importance.

In the past 10 or 15 years, as a result of prospecting for oil and diamonds in Yakutia, this region has seen a broad development of geologic survey and scientific research operations, whose scope has included the Vilyuy basin. Among these later works, the most important have been the investigations of A. A. Arsen'yev and V. A. Ivanova [1], which have provided a more detailed picture of A. G. Rzhonsnitskiy's scheme; on this basis (according to G. Ya. Krymgol'ts's identifications) the section has been divided into two layers with sand and conglomerate sediments (the Yemyaksinskian and the Ukugutskian), as well as strata containing Harpax (Middle Liassic), those containing Leda (Upper Liassic), a marine Aalenian layer with Ludwigia murchisonae Sow., and continental deposits of the undivided Middle and Upper Jurassic.

Of great value for the stratigraphy of the Jurassic deposits is also the paper by V. A. Vakhrameyev and Yu. M. Pushcharovskiy [3] on the adjacent areas, in which they have distinguished almost the same strata as along the Vilyuy River (containing Harpax in the Middle Liassic and Leda in the Upper Liassic).

Mention should also be made of a number of investigations by A. G. Kossovskaya [5], who was the first to note the typical mineral associations for the Middle and Upper Liassic deposits of the northwestern flank of the Vilyuy syneclise.

Beginning in 1950, regular geologic surveys on the scales of 1:1,000,000 and 1:200,000 have been made in the Vilyuy River basin by the All-Union Aerogeological Trust (with the participation of the Saratov V. N. I. I.). These organizations have done much important work in

[1]Novyye dannyye po stratigrafii Nizhneyurskikh morskikh otlozheniy r. Vilyuya, (pp. 88-98).

mapping the Jurassic deposits. Directly along the Vilyuy River, investigations were made by V. I. Kurlayev, B. K. Gortsuyev, and V. N. Bgatov, following either G. Ya. Krymgolts' or A. A. Arsen'yev's scheme for the subdivision of the Jurassic deposits. But the definite description of the Jurassic deposits of this area was made somewhat later, in the Explanatory Notes accompanying the Map Sheet R-50, prepared by R. E. Treylob, B. N. Leonov, and G. F. Lungersgauzen with the participation of B. P. Vysotskiy [13].

An important practical contribution to the working out of the unified stratigraphic schemes was made by the Conference on the Stratigraphy of Siberia, which was organized in 1956 in Leningrad. This conference took up the problem of the stratigraphic subdivision into stages of the Jurassic marine deposits of the Vilyuy syneclise and the Verkhoyansk basin, proposed by Z. V. Koshelkina [6, 7]. Along the Vilyuy River, this scheme recognized the Domerian, Toarskian and Aalenian stages among the marine deposits. The Aalenian stage is characterized by single specimens of the ammonite Ludwigia murchisonae Sow., from the collections made by A. G. Rzhonsnitskiy.

In 1958 was published the valuable synthesizing paper by V. A. Vakhrameyev [4], which, on the basis of the author's personal observations and the literature sources, brought together all the known information on the Jurassic stratigraphy of the Vilyuy River.

After the stratigraphic conference G. Ya. Krymgol'ts, in looking over A. G. Rzhonsnitskiy's collection, noted some inaccuracies in the identifications of the ammonites Ludwizia murchisonae Sow, and undertook a review of the paleontological remains.

At the same time, for the purpose of making additional collections of ammonites and finding outcrops of the strata from which A. G. Rzhonsnitskiy had collected the above-mentioned ammonites, the geologist N. N. Tazikhin visited the Vilyuy River in 1957. He did not, however, find any bedrock exposures of the strata containing Ludwigia murchisonae Sow. and collected only a few samples of ammonites from the talus of an outcrop that had already been described earlier by A. G. Rzhonsnitskiy.

In his monographic treatment of the ammonites collected by A. G. Rzhonsnitskiy and N. N. Tazikhin, G. Ya. Krymgol'ts reidentified the ammonites called Ludwigis murchisonae and placed them in a new genus Osperleioceras (O. viluiense). On the basis of the similarity of certain morphological criteria between these forms and the Western European species, he also made some important suggestions in regard to the Middle Toarskian age of the Vilyuy remains. His proposal radically changed the conception of the Viluyu section and cast doubt on the presence of Aalenian deposits in it, though they had been firmly established in the literature.

Nevertheless because A. G. Rzhonsnitskiy's discoveries of bedrock occurrences were never repeated, and because numerous remains of the same forms had been collected by N. N. Tazikhin from talus, it appeared that the essential problem of the age of the upper half of the marine deposits containing Leda was still far from settled. This supposition still required confirmation, and it remained necessary to find a satisfactory answer to the question of the age of the upper half of the Lower Jurassic marine deposits in the section.

The solution of this important problem is connected with the preparation and publication of the series of sheets of the State Geologic Map on the scale of 1:200,000, and with the correct interpretation of the geologic history of the Vilyuy syneclise. With this goal in mind, the present writer undertook her trip to the Vilyuy River in the spring of 1958.

The materials derived by this writer from the outcrops along the Vilyuy River now make it possible to present the stratigraphy of the section through the Jurassic marine deposits in considerable detail.

STRATIGRAPHIC OUTLINE

Lower Jurassic

The Domerskian stage (J_1d). The deposits of the Domerskian stage occur along the Vilyuy River in the area from the mouth of the Ulakhan-Dzhiyelligir River to Belesyusek-Aryytta Island and conformably overlie the sandstones and conglomerates of the Ukugutskaya formation.[2] They also merge gradually into the overlying deposits of the Toarskian stage.

The deposits of the Domerskian stage are represented by sands, sandstones, sandy shales and calcareous sandstones, which are frequently rich in plant material and in places even contain thin lenses of brown coal. Total thickness no more than 36 m.

On the basis of the paleontological features of the deposits, the Vilyuy section may be taken as the typical section through the Domerskian stage of the Vilyuy syneclise.

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[2]According to this writer's information, the geologic age of the Ukugutskian formation is Lower and partly Middle Liassic.

Below follows a bed-by-bed description of the deposits of the Domerskian stage, according to the individual outcrops which contain paleontological remains.[3]

In the section near the mouth of the Ulakhan-Dzhiyelligir River, on the right bank of the Vilyuy River, the exposures from bottom to top are:

J_1uk 1-7. Massive coarse-grained gray and yellowish-gray sandstone, highly micaceous. 25 m.

J_1d 8. Ochre-yellow sand with interbeds of dark gray clay and small lenses of coal. 0.1 m.

9. Dark-gray clay, very sandy, with layering a millimeter thick, noticeably rich in plant detritus and with small lenses of coal. In the lower half of this clay stratum the present writer has identified Pseudomonotis tiungensis Petr., Panope sp., and Nannobelus ex gr. janus Dum. 17 m.

10. Highly ferruginous, coarse-grained sand, with scattered pebbles of quartz and metamorphic and sedimentary rocks, and nodules of sideritized sand concretions and shell fragments of the genus Harpax. 0.15 - 0.2 m.

11. Gray, coarse-grained sand with small lenses of brown coal. 1 m.

12. Dark gray sand full of plant detritus. 0.08 m.

13. Gray coarse-grained sand. 1.6 m.

14. Massive gray sand, with three interlayers of loaf-shaped concretions of calcareous sandstone. The bottoms of the lenses of calcareous sandstones contain Amaltheus margaritatus Montf. The calcareous sandstones themselves contain Harpax terquemi Desl., H. aff. originalis Kosch., H. viluicus Kosch., Panope elongata Kosch., Pleuromya liasica Kosch. and Myophoria cf. batuobica Kosch. 3.79 m.

15. Fine-grained gray sandstone with interlayers of cross-bedded ferruginous sandstone. 1.9 m.

16. Dark gray shales, highly sandy, with occasional nodules of dark gray (bluish-gray on weathered surfaces) limestone, containing rare Pseudomonotis aff. tiungensis Petr. and Paltarpites argutus Buckm. 8.9 m.

17. Ochre-yellow gypsiferous sandstone. 1 m.

Jt 18. Dark gray shales with layers of sand and concretions of calcareous sandstone containing Leda acuminata (Golff.). 1.6 m.

In generalizing the material on these outcrops, it is important to note that the Domerskian stage is lithologically heterogeneous and is subdivided into three packets: a lower packet represented by sandy shales, a middle one composed of sandstones, and an upper packet of sandy shales.

Eastward, upstream along the Vilyuy River, some 2.5 km from Belesyusek-Aryytta Island (on the right bank), the upper half of the Domerskian section is exposed. This, from bottom to top, contains:

J_1d 1. Dark gray calcareous sandstone grading into coarse-grained sandstone with Harpax terquemi Desl., H. ex gr. laevigatus Orb., H. Viluicus Kosch., Pleuromya striatula Ag., Lenella tiungensis Kosch. In the uppermost part of the sandstones have been identified Pseudomonotis tiungensis Petr. and Tancredia kuznetsovi Petr. 3 m.

2. Fine-grained, dark gray sandstone, ferruginous in places. 2 m.

3. Yellowish-brown clayey sandstone. 1 m.

4. Gray fine-grained sandstone, with interlayers of sideritized sandstone concretions and nodules of calcareous sandstone. 0.7 m.

J_1t 5. Sandy light-yellow clay, with lenses of sandstone. 1 m.

The conclusion regarding the geologic age is based on the assemblage of fossils: the ammonites Amaltheus margaritatus Montf. and Paltarpites argutus Buckm. occurring above it, and also on the characteristic belemnites and bivalve molluscs. Among these, the ammonite Amaltheus margaritatus Montf. is extremely widespread and is known in almost all the main sections in Central and Northern Yakutia. The numerous and frequent occurrences of Amaltheus margaritatus in the Domerskian deposits of Yakutia make it possible to use this form for zonal stratigraphy. This is also the place to note that the presence of deposits below the Amaltheus margaritatus zone has thus far not been established in the territory of the Vilyuy syneclise. It is not impossible that these are continental facies of the Ukugutskian formation. Nor is there any paleontological basis for the presence of these deposits in Northern Siberia.

A comparison of the sections through the Domerskian stage on the Vilyuy River with those in the adjacent areas of the Markha and Tyung

[3]The cephalopods from the Vilyuy section were identified by the present author in consultation with G.Ya. Krymgol'ts and G.F. Lungersgauzen.

Rivers (Figure 1) reveals the fact that the latter are distinguished from the main section by their lithologic composition, by their somewhat different fossil contents and by their inconstant thicknesses. For example, along the

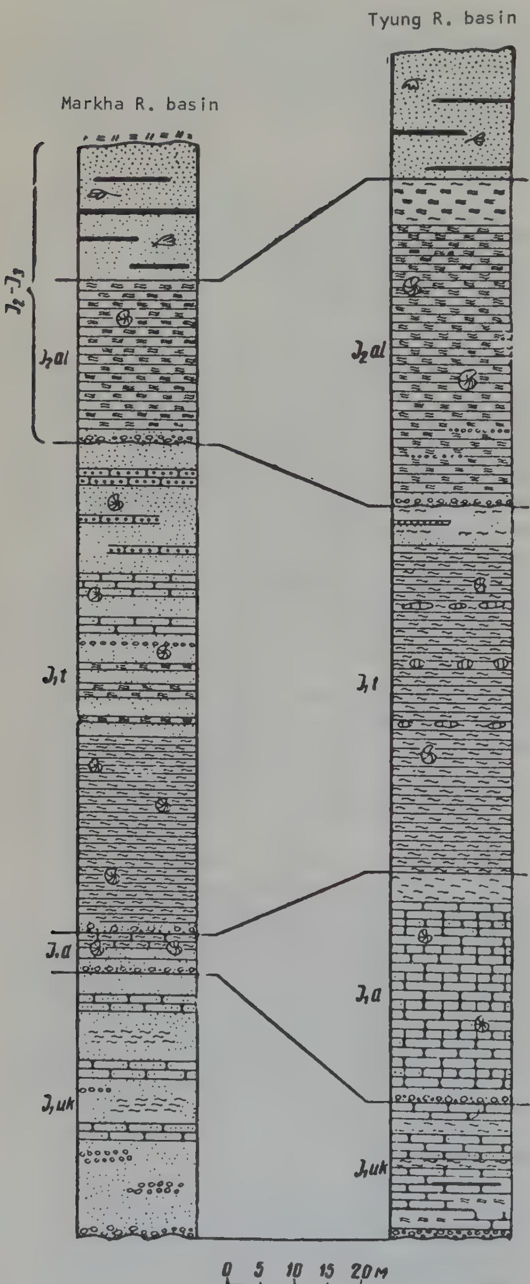

FIGURE 1. Correlation of the sections through the Jurassic marine deposits along the Markha and Tyung Rivers (drawing upon data from V.A. Vakhrameyev and Yu.M. Pushcharovskiy).

Markha River the deposits of the Domerskian stage lie sometimes upon the sandstones of the Ukugutskian formation with pebbles at the base of the formation, and sometimes on the Paleozoic rocks and are represented entirely by gray fine-grained sands and sandstones. The thickness of the deposits along the Markha River is small (according to G.I. Bushinskiy) — from 0.5 to 6 m. Completely different relationships obtain in the Tyung River section, where similar deposits in places also transgressively overlie the Paleozoic limestones. The latter section contains two packets in all: the lower represented by an alternation of polymictic sandstones and limestones, and the upper of sandy foliated shales. The thickness of the Domerskian stage along the Tyung River is evidently somewhat greater than along the Vilyuy River. Domerskian deposits have also been mentioned along the Batuobiya River by A. A. Arsen'yev [2].

The Toarskian stage (J_1t). Deposits of the Toarskian stage are widely known along the Vilyuy River. They are exposed in the area between the mouths of the Ulakhan-Dzhiyelligir and the Yulyuger Rivers. In all three outcrops these deposits are conformably underlain by the sandstones and shales of the Domerskian stage and are overlain (with conglomerates at the base) by coal-bearing deposits of Middle Jurassic age. The section through the Toarskian stage consists of a very homogeneous series of sandy, bluish-gray shales with frequent pecularly shaped lenticles or nodular elongated concretions of sandstones and pelitomorphic and sandy limestones, which as a rule are full of faunal remains. The visible thickness of the Toarskian deposits along the Vilyuy River is probably no more than 47 m. The present writer has studied the deposits of the Toarskian stage in three outcrops.

In the first outcrop, located near the mouth of the Ulakhan-Dzhiyelligir River, the lower half of the section is exposed. Here the following beds have been described, from bottom to top:

J_1d 17. Ochre-yellow gypsiferous sandstone. 1 m.

J_1t 18. Dark gray shales, with layers of ferruginous sand and concretions of calcareous sandstone containing Arctotis sp. and Leda acuminata (Goldf.). 1.6 m.

19. Dark gray shale with concretions of pelitomorphic limestone. 9 m.

20. Dark gray sandy shale with interlayers of concretions that are disposed at the following intervals from the bottom of the stratum: a) 4.38 m, dark gray argillaceous sandstone; b) 0.78 m, dark gray sandy limestone with Phacoides sp.; c) 1.14 m, dark gray calcareous sandstone with many Leda jacutica Petr.; d)

3.92 m, dark gray pelitomorphic limestone with Leda acuminata (Goldf.), L. jacutica Petr., Tancredia sp., Modiolus nitidula tiungensis Petr.; Mesoteuthis gracilis Hehl., M. ex gr. stimula Dum., and Septaliphoria sp.; e) 1.8 m, dark gray calcareous sandstone with rare Leda; f) 3.9 m, greenish-gray sandstone.

21. Greenish-gray sandy shale. 4 m.

22. Dark gray sandy shale with layers of sideritic concretions. 4 m.

Downstream along the Vilyuy River, about 2.5 km from Belesyusek-Aryytta Island, is the second outcrop of the deposits with which we are concerned here; this contains a fuller section of the beds with Leda.

This outcrop (first described in 1917 by A. G. Rzhonsnitskiy) extends for some distance but is less well exposed than the first. Upon the deposits of the Domerskian stage, from bottom to top, lie:

J_1t 5. Light-yellow sandy clay with small lenses of sandstone. 1 m.

6. Rusty brown sideritized sandstone. 0.1 to 0.15 m.

7. Dark-gray sandy shale with lenses of light-gray clay and siderite nodules, and also fragments of Arctotis sp. 4.3 m.

8. A pocket of lenticular concretions of dark gray argillaceous limestone with Dactylioceras gracile Simps. 0.2 to 0.3 m.

9. Dark gray sandy shale. 1.2 m.

10. Yellowish-gray sandy shale with lenses of clayey sandstone. 5.8 m.

11. Dark yellow clay with large spherical concretions of sandy limestone. 2 m.

12. Rusty brown to reddish-cinnamon sandstone, coarse-grained, with small layers of gray sandstone and sandy limestone. 1 m.

13. Light yellow sandy clay. 0.5 m.

14. Rusty brown, medium-grained sandstone. 0.2 m.

15. Light yellow clay with pockets of calcareous sandstone concretions. 0.8 m.

16. Dark gray sandy shale with small lenses of dark cinnamon-colored clay.

17. Dark gray sandy shale with lenses of dark cinnamon-colored argillites. 3 m.

18. Dark gray sandy shale with small lenses of limestones, sandstones, clays and coquina. The order of occurrence of the fossils, with the thicknesses of the layers in which they are found, is the following (from bottom to top): a) 3.6 m, a layer of pelitomorphic limestone concretions with Leda jacutica Petr., L. acuminata (Goldf.), (Mytiloides) oviformis Khud., Mesoteuthis ex. gr. oxycona Hehl.; b) 1.8 m, a layer of shell limestone with Leda jacutica Petr., L. ex gr. acuminata (Goldf.), Tancredia sp., Mesoteuthis ex gr. stimula Dum., M. gracilis Hehl.; c) 1.1 m, a layer of shell limestone with Leda jacutica Petr., L. ex gr. acuminata (Goldf.), Tancredia sp., Osperleioceras viluiense Krim.; d) 1.6 m, a non-fossiliferous sandy limestone concretions; e) 1.8 m, concretions of sandy limestone and small banks with Leda acuminata (Goldf.), L. jacutica Petr., Mesoteuthis ex gr. stimula Dum.

19. Dark gray sandy shale with interlayers of clayey sandstone and small banks with Leda ex gr. acuminata (Golf.), Tancredia ex gr. stubendorffii Schm. 2.8 m.

20. A pocket of calcareous sandstone concretions with Leda viluiensis Kosch., Modiolus viluiensis Khud., Liostrea aff. acuminata Sow., Tancredia aff. stubendorffi Schm. 0.2 m.

21. Dark gray shale with a predominance of interlayers of dark cinnamon colored argillites and concretions of sandy limestones with Leda liluiensis Kosch., Tancredia aff. stubendorffi Schm., Lima sp. and plant traces not well enough preserved for identification. 8.8 m.

22. Sandy clay, highly ferruginous, with lenses of coal and concretions of sideritized sandstone with Arctotis ex gr. marchaensis Petr. and (Mytiloides) marchaensis Petr. 0.5 m.

J_2jk 23. Gravelite with large scattered pebbles. This layer has a lenticular structure and does not extend far laterally. 0.2 to 0.65 m.

24. Light gray and yellowish-gray sands with small lenses of gravelite and coal (Yakutshaya formation). 15 m.

The total thickness of the deposits in the Toarskian stage in this outcrop is no more than 46 to 47 m.

Downstream on the Vilyuy River, about 2 km below the mouth of the Yulyuger River, is the third and last comparatively large outcrop of deposits of the Toarskian stage (Figure 2). The rocks of this outcrop correspond stratigraphically to the upper half of the section of the preceding outcrop. Here, from bottom to top, are:

J_1t 1. Light yellow clays. 0.7 m.

2. Greenish-gray siltstone with Tancredia

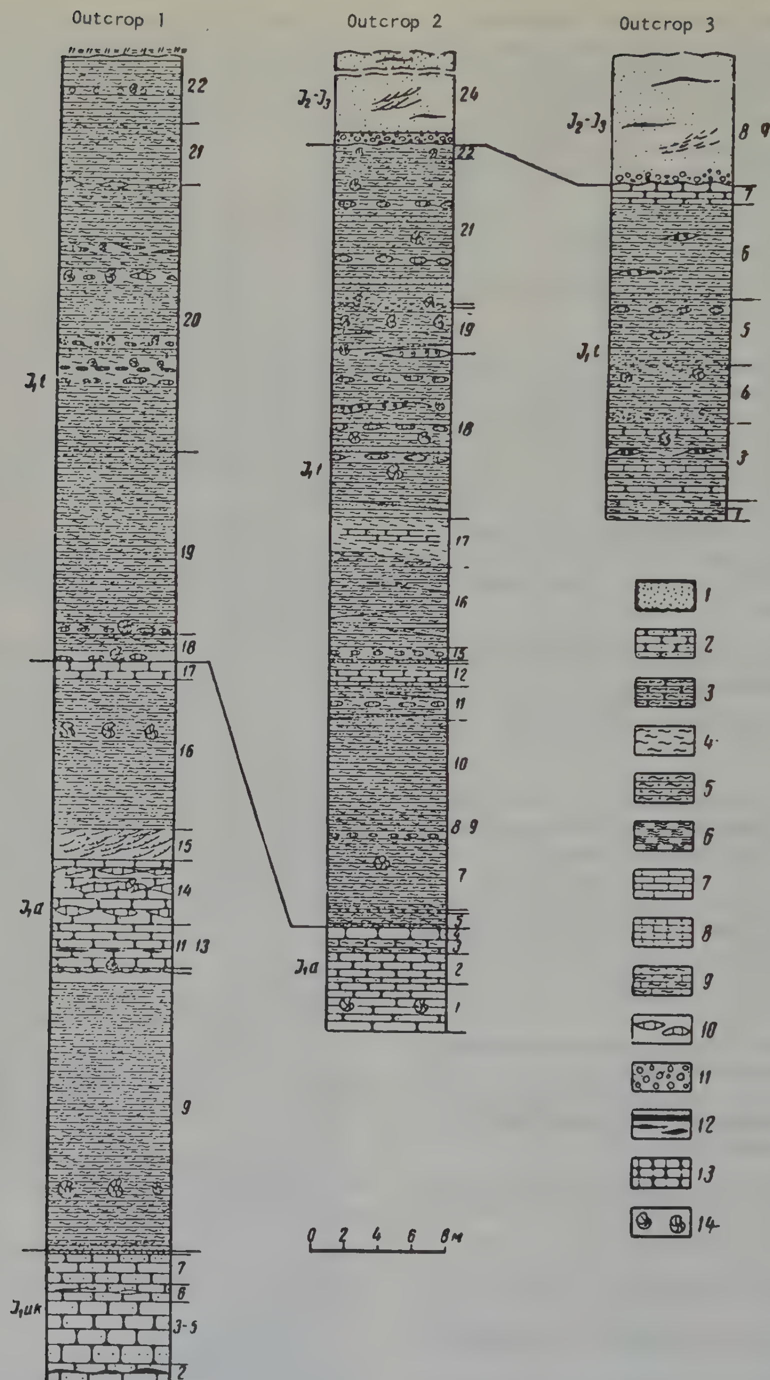

FIGURE 2. Correlation of the sections through the Lower Jurassic deposits along the Vilyuy River (Outcrops 1, 2 and 3 - after data of the present writer):

1 - sand; 2 - sandstone; 3 - clayey sandstone; 4 - clay; 5 - sandy clay; 6 - siltstone; 7 - limestone; 8 - sandy limestone, 9 - clayey limestone; 10 - limestone lenses and concretions; 11 - gravel; 12 - coal lenses; 13 - calcareous sandstone; 14 - paleontological remains.

sp., Leda jacutica Petr., L. acuminate (Goldf.). 0.5 m.

3. Greenish-gray sandy shale with layers of yellowish-gray siltstones and sandstones containing Leda sp. and Tancredia sp. 4.7 m.

4. Greenish-gray clay, sandy, with small lenses of cinnamon-colored clay and Leda jacutica Petr., Mesoteuthis ex gr. oxycona Hehl. 3.5 m.

5. Ochre-yellow clay with concretions of sandy limestone. 3.9 m.

6. Dark gray, sandy shale with interbeds of brown shale and banks containing Tancredia aff. stubendorffi Schm., Tancredia sp., Natica sp., Mesoteuthis aff. stimula Dum. and vertebrae of Eretmosaurus ex gr. rzonsnickii Menner.[4] 5.7 m.

7. Rusty brown, fine-grained sandstone with concretions of siderite. 0.5 to 1 m.

J_2jk 8. Gravelite, strongly cemented, with occasional large pebbles. 0.2 to 0.65 m.

9. Coarse-grained gray sand, in places ferruginous, with small lenses of coal. 5 to 7 m.

The geologic age of the deposits with widespread and abundant Leda fossils is also determined by a complex of forms: the ammonites Dactylioceras gracile Simps. and Osperleioceras viluiense Krimh. and the belemnites Mesoteuthis gracilis Hehl., M. stimula Dum. and M. oxycona Hehl., which are typical of the Toarskian stage both in Western Europe and in the U. S. S. R. In addition to these, the typically Toarskian bivalve molluscs Leda jacutica Petr. and L. acuminata (Goldf.) are also found. In view of the occurrence of Osperleioceras almost at the very top of the exposure of the marine section (in the bedrock outcrop) and of the similarity in morphology between this ammonite and the Western European species, it can be said that along the Vilyuy River the section described here does not go beyond the limits of the Middle Toarskian. Aalenian deposits are absent. The ammonite remains suggest a subdivision of the Vilyuy River section into two parts: the lower with Dactylioceras[5] and the upper with Osperleioceras. In this regard it may be noted that the Dactylioceras beds are quite widespread. They occur, besides the Vilyuy River, along the Markha and Tyung Rivers and even north of this territory; they apparently have some zonal significance.

Lithologically the Vilyuy River section differs somewhat from the very same sections along the Markha and Tyung Rivers. In the latter the deposits of the Toarskian stage are divided into two packets, the lower one argillaceous and the upper sandy. Along the Tyung River the top of the upper, sandy packet also contains foliaceous clay. In addition, along the Markha and Tyung Rivers one observes a regular increase in the thickness of the Toarskian deposits from 56 to 65 m. Below are the main assemblages of guide fossils for the Lower Jurassic marine deposits on the northwestern margin of the Vilyuy syneclise (the Vilyuy, Markha, Tyung and Batuobiya Rivers).

I. Domerskian stage (J_1d). Ammonites: 1) Amaltheus margaritatus Montf., 2) Paltarpites argutus Buckm. Belemnites: 3) Nannobelus janus Dum. Bivalves: 4) Harpax terquemi Desl. 5) Harpax viluicus Hosch., 6) Harpax spinosus Sow., 7) Myophoria batuobica Kosch., 8) Pseudomonotia tiugensis Petr., 9) Tancredia kuznetsovi Petr., 10) Pleuromya liasica Kosch., 11) Lenella tiugensis Kosch.

II. Toarskian stage (J_1t). Ammonites: 1) Dactylioceras gracile Simps., 2) Dactylioceras athleticum Simps.,[6] 3) Dactylioceras suntarenae Krimh., 4) Osperleioceras viluienae Krimh., Belemnites: 5) Mesoteuthis oxycona Hehl., 6) Mesoteuthis stimula Dum. Bivalves: 7) Arctotis marchaensis Petr., 8) Leda acuminata (Goldf.), 9) Leda jacutica Petr.

Among the forms shown in the table, along with the typical fossil complexes, this writer has identified a rare ammonite. Since this ammonite was first found in the Jurassic deposits of the U. S. S. R., it seems suitable to present a description of it at this point.

DESCRIPTION OF AMMONITE REMAINS

Paltarpites Argutus Buckman, 1923,
Figure 3 [1, 2]

1927. Paltarpites argutus. Buckmann. — Type ammonites, t. 393. Figure 1, 2; Ser. 37.

Description. The shell is of medium size, disc-shaped, highly compressed laterally, and evolute. It has a broad umbilicus and faint chamber walls. The siphonal side is sharply carinate, and the keel is not clearly

[4]Vertebrae of Eretmosaurus were identified and described from A.G. Rzhonsnitskiy's collection by V.V. Menner (Outcrop 2). The Eretmosaurus specimens from Outcrop 3 were collected by the present writer.

[5]From the talus of Outcrop 2 N.N. Tazikhin has collected Dactylioceras gracile Simps., D. suntarense Krimh, and Osperleioceras viluiense Krimh.

[6]The guide fossil assemblages were distinsuished both from personal observations and from the data of V.A. Vakhrameyev, G.I. Bushinskiy, Yu.M. Pushcharovskiy and G.F. Lungersgauzen.

Distribution of the Paleontological Remains in the Sections along the Vilyuy, Markha, Tyung and Bol'shoy Batuobiya Rivers. Domerskian Stage (J_1d)

No.	Name of organism	Vilyuy R.	Markha R.	Tyung R.	Batuobiya R.	No.	Name of organism	Vilyuy R.	Markha R.	Tyung R.	Batuobiya R.
1	Amaltheus margaritatus Montf.	+	+?	+	—	12	Tancredia kuznetsovi Petr.	+	—	—	—
2	Paltarpites argutus Buckm.	+	—	—	—	13	Pleuromya liasica Kosch.	+	+	+	+
3	Nannobelus janus Dum.	+	+	—	—	14	Pl. striatula Ag.	+	—	—	—
4	Harpax terquemi Desl.	+	+	—	—	15	Pl. galathea Ag.	—	+	—	—
5	H. laevigatus Orb.	+	—	+	—	16	Panope lahuseni Kasch.	—	—	+	—
6	H. viluicus Kosch.	+	—	—	—	17	P. elongata Kosch.	—	—	+	—
7	H. spinosus Sow	—	+	+	—	18	Solen liasicus Opp.	+	—	—	—
8	Myophoria batuobica Kosch.	+	+	+	+	19	Modiolus nitidula tiungensis Petr.	+	—	—	+
9	Pseudomonotis tiungensis Petr.	+	+	+	+	20	Turbo khudjaevi Petr.	—	+	—	—
10	Ps. sparsicosta Petr.	—	—	+	—	21	Pleurotomaria aff. singularis Sieb.	—	+	—	—
11	Lenella tiungensis Kosch.	+	+	+	—						

Toarskian Stage (J_1t)

No.	Name of organism	Vilyuy R.	Markha R.	Tyung R.	No.	Name of organism	Vilyuy R.	Markha R.	Tyung R.
1	Dactylioceras gracile Simps.	+	+	+	9	Nannobelus pavlovi Krimh.	—	—	+
2	D. athleticum Simps.	—	—	+	10	Orthoceras sp.	—	—	+
3	D. suntarense Krimh.	+	—	—	11	Leda jacutica Petr.	+	+	+
4	Osperleioceras viluiense Krimh.	+	—	—	12	L. acuminata (Goldf.)	+	+	+
					13	Arctotis marchaensis Petr.	—	+	+
5	Mesoteuthis oxycona Hehl.	+	—	+	14	(Mytiloides) marchaensis Petr.	+	+	—
6	M. gracilis Hehl.	+	—	—					
7	M. stimula Dum.	+	—	+	15	(M.) oviformis Khud.	+	+	—
8	Passaloteuthis sp.	+	—	—	16	Modiolus numismalis Opp.	—	+	—

distinguished from the lateral surfaces of the shell. In the fossil impression one may trace wavy, curving thin ribs. The amount of these ribs in the adult whorl (per 1/4 d) is about 30. The ribs merge without a break into the keel, giving the latter a beaded appearance. The young whorls as a rule have bundles with different successions of ribs (Figure 2).

Dimensions. The greatest diameter is 89 mm; the height of the whorl is 34 mm; the diameter of the umbilicus is 33 mm?; the ratio of the whorl height to the shell diameter is 0. 38.

Remarks. The present species may be identified with the form first described by S. Buckman from the top of the Domerskian stage [14] in England (Vol. 363; Figures 1, 2; Series 37; 1923). It differs from Paltarpites paltus Buckm. (Vol. 362; Figures 1, 2; 1922) in having thinner and more numerous ribs on the earlier whorls and in its smaller shell diameter.

Distribution. The upper part of the section through the Domerskian stage of England (above the Amaltheus margaritatus zone).

Location. Central Yakutia, Vilyuy River, in the vicinity of the mouth of the Ulakhan-Dzhiyelligir River (above the beds containing Amaltheur margaritatus Montf.)

Collected by the present author (Museum of the All-Union Aerogeological Trust).

FIGURE 3. Imprint of the ammonite Paltarpites argutus Burkm. from the Vilyuy River (the area near the mouth of the Ulakhan-Dzhiyelligir River). Domerskian stage (top). Collected by Z.V. Koshelkina. Two-thirds natural size.

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BRIEF COMMUNICATIONS

SYNTHETIC INTRODUCTION OF ARGON INTO MICA AT HIGH PRESSURES AND TEMPERATURES[1]

by

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The widespread use of the potassium-argon method in geochronology has aroused considerable interest in the problem of the retention of radiogenic argon in potassium-containing minerals.

The experimental investigation of the actual process of the loss of radiogenic argon under conditions close to those obtaining in nature involves great difficulties. The separation of argon from feldspars, and all the more from micas, at temperatures of 500°C or lower takes place extremely slowly (it is for this very reason that the potassium-containing minerals are so suitable for the determination of absolute age). Thus in the overwhelming majority of cases the loss of radiogenic argon is studied at higher temperatures and the resulting laws or formulas are extrapolated into the region of lower temperatures.

The present writers have found it feasible to approach the study of this problem from the opposite side: to investigate the reverse process, the introduction of argon from the gaseous phase into the mineral.

By using high pressures, it has been possible to produce a considerable increase in the rate of diffusion of argon into the mineral, without raising the temperature. By measuring the mean content of argon introduced into the mineral at two values of the diffusion criterion Dt/a^2 (t is the duration of the introduction and a is the particle size of the mineral), one may calculate the diffusion coefficient D. (The results of this investigation can, of course, be used as a measure of the loss of radiogenic argon from the mineral only if it can be shown experimentally that the mechanism in both processes is identical. The introduction of argon in this experiment should therefore be carried out under conditions that will ensure the greatest possible retention of the basic properties of the mineral.)

There is also another possible approach, using high pressures to study the migration of argon in the mineral — this amounts to a synthesis of the mineral under high argon pressure, resulting in a mineral with a much higher argon content than under natural conditions and in naturally formed minerals. With a high content of argon, it is much easier to study the process of segregation of this element (using the same methods as in the case of radiogenic argon).

At the first, preliminary stage of this work, which is set forth in he present article, the authors have set themselves the task of finding answers to two questions:

1. Is argon introduced into minerals under high pressures? (Muscovite was taken as the material for this study).

2. What is the nature (strength) of the bond that retains the introduced argon within the mineral?

Experiments in the introduction of argon into muscovite at pressures of 3,000 to 5,000 atm and temperatures of 750 ti 850°C were carried out as follows: natural muscovite (0.5 g), broken down to particles of 0.2 mm size, and also muscovite in the form of platelets 4 x 30 mm in dimensions, were placed in an open platinum ampoule inside a high-pressure bomb, and were held for several hours at high temperature under argon pressure. The content of argon in the muscovite after the experiment was determined with a mass spectrometer by the isotope dilution method. The results are given in Table 1.

[1]Iskusstvennoye vnedreniye argona v slyudu pri vysokikh davleniyakh i temperaturakh, (pp. 99-100).

Table 1

Experiment No.	Material studied	Argon pressure, atm	Temperature degrees	Duration, hours	Content of argon in muscovite after experiment in nmm^3/g
5	Macerated muscovite	2800	740	3	67
12	Same	3000	760	7	68
13	"	4400	825	10.5	89.5
15	Muscovite platelets	5000	860	5	9

Table 1 shows that, in the first place, the muscovite that has undergone high argon pressure absorbs a considerable amount of argon, many times exceeding the content of radiogenic argon in the original muscovite, which was some 0.12 nmm^3/g; and, in the second place,

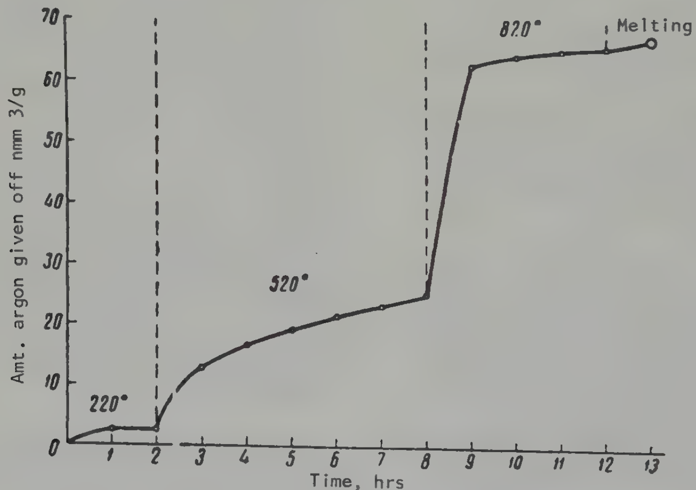
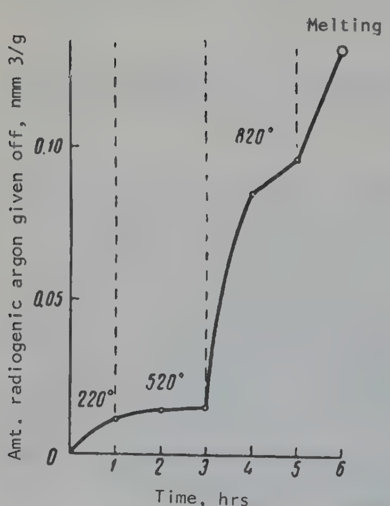


FIGURE 1. Separation of argon absorbed at high pressures from muscovite at various temperatures.



that the degree to which the muscovite has been macerated apparently plays an important role in the process of absorption of argon at high pressures, inasmuch as there is a great difference in the amount of argon absorbed by the

FIGURE 2. Separation of radiogenic argon from muscovite at various temperatures.

macerated muscovite (70 to 90 nmm^3/g) as compared to that absorbed by the muscovite platelets (9 nmm^3/g).

To clarify the process of absorption of the argon, the absorbed argon was separated out from the mica at various temperatures: 220°,

20°, and 820°C. Figure 1 shows one of the curves characterizing the separation of argon from the macerated muscovite, which has been held for 7 hours at the temperature of 720° and the argon pressure of 3000 atm (experiment 2).

As may be seen from Figure 1, about 60% of the argon absorbed at high argon pressure is separated from the muscovite at temperatures higher than 520°. The greater part of the radiogenic argon is also lost from the original muscovite at temperatures higher than 20° (Figure 2).

These data suggest that much of the argon absorbed into the mica at high pressures is strongly held by the mineral and evidently enters into its lattice, since in the case of mere surface adsorption one would have expected all the adsorbed argon to be separated from the mineral at low temperatures.

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THE CONTENTS OF ZIRCONIUM AND HAFNIUM IN THORTVEITITE[2]

by

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In the scandium mineral, thortveitite, that was discovered by J. Schetelig in the granite pegmatites of Southern Norway, zirconium and hafnium were originally not determined [9, 10]. Later on A. Lacroix [5] cited a content of 8.4% ZrO_2 in the thortveitite from the pegmatites from the Island of Madagascar (Befanamo district) — the average of three analyses, with a minimum of 7.4% and a maximum of 9.9% ZrO_2. The discovery of such high contents of zirconium made it seem advisable to distinguish an independent zirconium variety of thortveitite — befanamite [1, 2]. But shortly after the discovery of the Madagascar thortveitite, J. Schetelig pointed out that it was identical to the Norwegian thortveitite [3]; according to his data, the ZrO_2 content in the thortveitite from Norway is 5% (analysis by Zigban-Hading).

Somewhat later G. Hevesy and V. T. Jantzen presented the results of X-ray spectrum studies confirming the presence of ZrO_2 in the thortveitite of Norway and Madagascar, but in much smaller amounts than indicated earlier by A. Lacroix and J. Schetelig. Moreover, in addition to the ZrO_2, the presence of HfO_2 was established in both thortveitites (in % amounts; see Table 1).

The higher contents of zirconium and hafnium as determined earlier are apparently to be explained by the fact that instead of ZrO_2, the sum of the oxides of zirconium and hafnium was given (hafnium was discovered only in 1923). There may also have been some inaccuracy in the chemical analysis.

Nevertheless, in spite of the fact that the presence of zirconium (and hafnium) has been established in the thortveitites from both Norway and Madagascar, until recently the literature has continued to mention the zirconium variety of thortveitite — befanamite — only in the case of Madagascar.

The present writers have made spectrum analyses of thortveitites from Norway and Madagascar. The specimens were obtained

Table 1

Area	ZrO_2	HfO_2	$HfO_2:ZrO_2$	Investigator
Unneland, Norway	0.8	1.1	1.37	G. Hevesy and V. T. Jantzen, 1924
Iveland, Norway	1.2	2.0	1.67	Same
Befanamo, Madagascar	2.2	1.8	0.82	"
" "	2.0	3.2	1.60	"
" "	1.3	1.0	0.77	G. Hevesy, 1925
Iveland, Norway	2.0	0.5	0.25	Same

from the Mineralogical Museum of the Academy of Sciences of the U. S. S. R. The material to be investigated was carefully picked out under the binocular microscope. The amounts of ZrO_2 and HfO_2 were determined by the spectrum method on three standards. Common to the

[2] O soderzhanii tsirkoniya i gafniya tortveytite, (pp. 101-103).

Table 2

Area	ZrO_2	HfO_2	$HfO_2:ZrO_2$	Y_2O_3	Investigator
Iveland, Norway	2.0	0.8	0.4	>2	Present authors
Befanamo, Madagascar	1.1	1.0	0.9	>2	"
Iveland, Norway	1.2	0.6	0.5	2.2	A. Levinson, R. Borup, 1960
" "	1.9	1.7	0.9	4.4	" "
Iveland, Norway	1.4	1.4	1.0	4.3	" "

standards was a mixture of 40% scandium oxide and 60% silicon dioxide. The spectra were photographed with an ISP-28 spectrograph with a three-lens condenser. A weighed sample of 2 mg of thortveitite was mixed with the same amount of carbon and placed in the crater of the machined carbon electrode (the anode). The crater diameter was 2 mm and its depth 3 mm.

A continuous-current arc was ignited between the vertically placed electrodes; the current was increased over a duration of 30 sec from 5 amp to 15 amp, and at 15 amp during the succeeding minute the contents of the lower electrode were completely vaporized. Calibrated curves were prepared for the coordinate (s; log C), and the II 2647. 29 and II 2567. 64 Å lines were used. The relative error for a single determination was approximately 20%.

In the analyzed specimens, the presence of zirconium and hafnium were determined in the amounts of 1 to 2% (Table 2). At the same time similar results were obtained, independently of the present authors, by A. Levinson and R. Borup [7].[3]

In addition, the present writers established the presence of a noticeable amount of yttrium, as well as iron, magnesium, manganese, aluminum, ytterbium and thorium, in both samples. The total contents of ZrO_2 and HfO_2 varied from 2. 1 to 2. 8% according to these writers' data, from 1. 9 to 4% according to G. Hevesy and R. Borup. Thus zirconium and hafnium occur in quite similar quantities in the thortveitite of Norway and Madagascar, so that there is no basis for setting aside the Madagascar thortveitite as an independent variety, to be called befanamite.

The thortveities of Norway and Madagascar also do not differ in their other properties. Their optical constants are very similar:

Thortveitite	α	β	γ	$\alpha - \gamma$	Investigator
Norwegian	1.756	1.793	1.809	0.053	J. Schetelig, 1922
"	1.751	1.789	1.803	0.052	J. Glass, 1942
Madagascar	1.750	—	1.800	0.050	A. Lacroix, 1923

Very similar X-ray powdergrams were also obtained from both kinds of thortveities in the X-ray structure laboratory of the Institute of Mineralogy, Geochemistry and Crystallography of Rare Elements of the Academy of Sciences of the U. S. S. R. (Table 3).

The paragenetic associations in which the thortveitites occur in the granite pegmatites of Norway and Madagascar are also much alike: the Norwegian thortveitite is associated with euxinite, ilmenorutile, monazite, beryl and biotite, and the Madagascar thortveitite with monazite, fergusonite, ilmenorutile, beryl, magnetite and muscovite.

In turning to the problem of the form in which the zirconium and hafnium occur in throtveitite, one must take note of their similar ionic radii (r_i) and the similar values of their electrical negativity (E), all resembling the respective constants for scandium:

[3]The article by A. Levinson and R. Borup [7] appeared in print after the present article had already been submitted to the editors of the journal "Izvestiya Akad. Nauk S.S.S.R., Seriya Geologicheskaya".

	Sc	Zr	Hf
Charge	3	4	4
Ionic radius (Å)	0. 83	0. 82	0. 82
Coordination number	6	6	6
Electrical negativity (kcal/g/atom)	200	200	(180)

From the similarity in these constants, it appears that zirconium and hafnium replace the scandium in the lattice structure of thortveitite. The charge is compensated by the aluminum, which according to the spectrum analysis data is present in both the specimens of thortveitite investigated. The presence of Al_2O_3 in the thortveitite of Norway was also established chemically by J. Marble [8], in the amount of 95%, and by Bauland and Urben (citation 6) in the Madagascar thortveitite in the amount of 3%.

The replacement of scandium by zirconium and hafnium may be schematically represented as follows:

$$Zr^{4+}, Hf^{4+} + Al^{3+} \quad Sc^{3+} + Si^{4+}.$$

Careful winnowing of the material investigated by the present writers under the binocular microscope has eliminated the possibility of the presence of any Zr and Hf as coarse mechanical admixtures. Hevertheless it must be admitted that a portion of these rare elements is contained in thortveitite in the form of microinclusions in zircon.

In this light it is interesting to note a quite recently discovered zircon, intergrown with a crystal of Norwegian thortveitite. This zircon contained about 22 to 24% HfO_2, with a Hf/Zr ratio of 0. 6. Such a very unusual ratio, as well as the close association between the zircon and the thortveitite, testify to the specific character of the melt-solution from which these two minerals were formed. Apparently at the stage of separation of these minerals, the melt-solution was enriched in hafnium (relative to zirconium), and this enrichment was reflected in the composition of both the zircon and the thortveitite.

Thus these investigations, along with the results obtained by other authors, show that zirconium (and hafnium) is contained in like quantities in the thortveitite from both Madagascar and Norway. As noted above, the

Table 3

X-ray Powdergrams for Thortveitite

Line No.	Thortveitite		Thortveitite, Madagascar		Line No.	Thortveitite		Thortveitite, Madagascar	
	I	d (kx)	I	d (kx)		I	d (kx)	I	d (kx)
1	1	5.16	—	—	19	1,5	1.373	1.5	1.366
2	—	—	1	3.452	20	1	1.328	1.5	1.323
3	10	3.14	10	3.110	21	<1	1.274	1	1.275
4	3.5	2.94	1.5	2.923	22	—	—	1	1.193
5	2	2.57	1	2.600	23	<1	1.166	1	1.164
6	—	—	1	2.538	24	1	1.087	2	1.084
7	2	2.18	2	2.166	25	<1	1.060	1.5	1.058
8	1	2.10	1	2.076	26	1	1.035	1.5	1.033
9	1.5	2.06	2	2.040	27	<1	1.010	—	—
10	1.5	1.885	1.5	1.868	28	<1	1.001	—	—
11	—	—	1.5	1.808	29	2	0.962	—	—
12	1.5	1.694	1.5	1.692	30	<1	0.947	—	—
13	3	1.643	2	1.637	31	<1	0.913	—	—
14	1.5	1.598	1	1.585	32	1	0.888	—	—
15	2	1.517	1.5	1.509	33	<1	0.879	—	—
16	<1	1.497	1	1.491	34	<1	0.815	—	—
17	<1	1.469	—	—	35	<1	0.809	—	—
18	1	1.421	1.5	1.416	—	—	—	—	—

Conditions of experiment:

1) K_a - Cu, Ni filter, 2R = 57.3 2r = 0.6 mm

2) K_a - Fe 2R = 57.3 2r = 0.6 mm

Analyst N.G. Bataliyeva

nearness of the crystallochemical constants of scandium and zirconium facilitates the introduction of zirconium into the composition of this mineral. None the less, the literature contains results of chemical analyses of thortveitite without zirconium and hafnium [8, 10]. It may be that in these analyses, in which the ZrO_2 and HfO_2 are absent, there was an error in their determination (or else that no tests were made for ZrO_2 and HfO_2 in general). If future examples of numerous other samples of thortveitite confirms the fact of the constant presence of zirconium, in small and approximately the same quantities without altering the structure of the mineral, serious doubt will be cast on the validity of distinguishing an independent zirconium variety — befanamite.

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THE MAIN FEATURES OF THE STRUCTURE OF THE WESTERN SIBERIAN PLATFORM[4]

by

V. N. Sobolevskaya

In 1959, at the Institute of Geology of the Academy of Sciences of the U. S. S. R., this writer completed the preparation of a tectonic map of the folded basement of the Western Siberian platform, as well as a structural map of the platform mantle along the surface of the Campanian-Maestrichtian deposits in the Transuralian part of the region.

In constrast to the Tectonic Map of the U. S. S. R. published in 1956 [3], this map shows not only the reworked and corrected structure contour lines of the surface of the basement complex, but also its division into tectonic regions.

In a brief article it is not possible to present this map and a full explanation of it, and this has been done in a special paper being prepared for printing; it is all the more impossible within the scope of an article to present a review of all the existing conceptions of the tectonic structure of this extensive territory. Such surveys may be found in a number of published works [1] and elsewhere.

The tectonic sketch map prepared by this writer, showing the structure of the surface of the platform's folded basement in structure contour lines, along with the principal associated structural elements, and the established and supposed boundaries of the present distribution of the Valanginian-Jurassic and the Rhaetian-Liassic deposits (see accompanying map), is set forth here with the single purpose of calling still greater attention to the tectonic structures in directing prospecting operations for deposits of various economic minerals.

At the present time, according to the most recent information ("Pravda", April 30, 1961), it may be said that the enormous efforts made by the geologists and geophysicists of Western Siberia have been rewarded with well-deserved success. Drilling operations in the Shaim area and the area of the Mulym'ya River in Northern Transuralia have, finally, shown the existence not only of gas, but also of oil within Western Siberia. It would be difficult to overestimate the importance of this fact for the U. S. S. R.'s national economy; nevertheless much still depends on the direction of exploration and prospecting work, which is in turn to a considerable

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[4]Osnovnyye cherty struktury Zapadno-Sibirskoy plity, (pp. 104-105).

degree determined by the existing conceptions of the structure of the region. These concepts should facilitate the discovery of regular relationships between the locally revealed gas- and oil-bearing structures and the chief tectonic elements, resulting in scientific prognoses and proper direction of further prospecting operations.

The Mulym'ya-Shaim oil-bearing region is located, according to the attached structural map, within a subequatorial zone of step-wise subsidence of the Epihercynian basement of the Western Siberian platform (map). The various tectonic zones distinguishable here were drawn into the subsidence along a system of faults running both parallel to the trend of folding of the basement rocks and transverse to it, cutting across the trend. The movements along these faults in most cases continued without interruption, or were renewed at certain intervals throughout the entire time of accumulation of the sediments composing the upper structural stage, forming a step in the basement and thus exerting a decisive influence on the formation and development of the structures in the platform mantle.

The existence of step-like depressions in the relief of the platform's basement has been established in a number of places by geologic and geophysical methods. The steps in the folded basement are reflected in the mantle by flexural bends of the strata, with relatively steep angles of dip at depth, the dips becoming smaller upward in the section, in the younger beds. It is suggested that these steps mark the boundaries between the deposits of different stratigraphic levels and different facies zones, as well as changes in thicknesses. For example, the Shaim disjunctive zone is, evidently, associated with the southwestern and western border of the area of Valanginian-Jurassic deposits.

In addition, the attached sketch map attempts to show that the disposition of most of the local structures found within the Lowland is not random, but is subordinated to a definite governing principle, which is to be found in their undoubted connection with the faults in the basement and the flexural folds of the mantle corresponding to these faults.

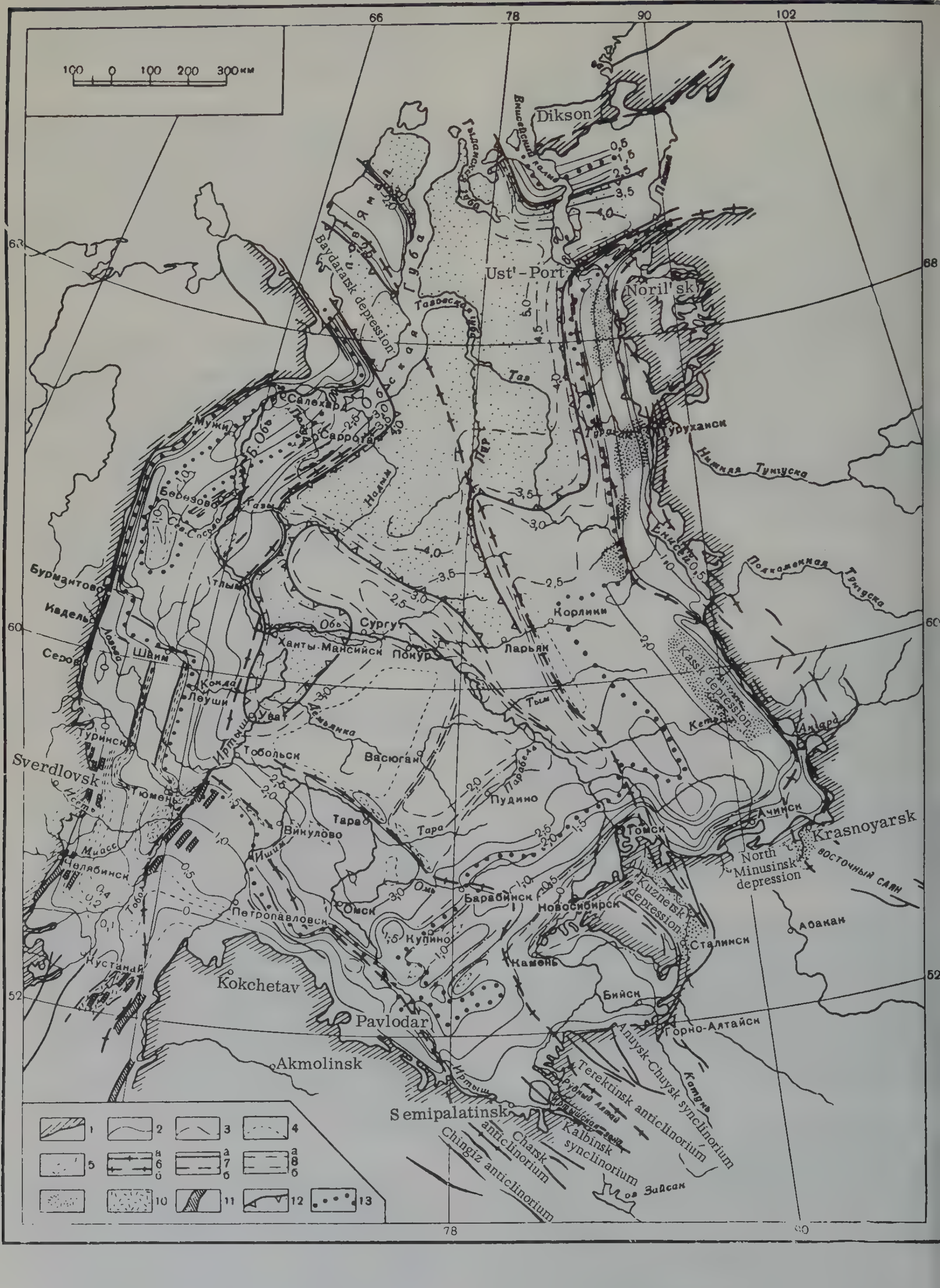
From what has been said, it follows that the principal tectonic lineaments of the basement determine the formation not only of the major first-order structures of the mantle, which give the whole territory its general tectonic aspect, but also the local structures that might serve as traps for oil and gas. Moreover one still cannot entirely dismiss the possibility that the faults may here have acted as channels by which the oil has risen from deeper strata. This concept deserves attention in this case, since in the Berezovka and Shaim districts, which are already known to be oil-bearing, there are Valanginian-Jurassic deposits of no great thickness directly overlying the folded basement. It is still an open question as to which are the formations in which the oil of Western Siberia originated.

Thus according to the conceptions set forth here, the Mulym'ya-Shaim oil is associated with a tectonic zone that cuts across the trend of the Urals, and lies in the vicinity of one of the meridional structural steps that run parallel to this trend, as shown on the attached structural map, and also on the map published in 1958, showing the structures of the basement of the Transuralian part of the Lowland [2].

The position of the above-mentioned drill holes, which have struck oil, calls attention to the zones of step-like depressions in the buried folded basement of the Western Siberian Lowland. One must keep in mind, however, that on the attached map they are still located by extremely few direct or indirect data. It goes without saying that much effort is still required to support these data and find similar new information, particularly on areas in which the basement occurs at depths of more than 1000 m; it is in these areas that one can expect to find the Valanginian-Jurassic deposits that have already yielded oil, under conditions favorable to their preservation. This, of course, does not exclude the possibility of finding economic concentrations of oil in the older and deeper strata of the mantle, which are quite likely to be present in the central and especially the northern parts of the Lowland.

There is also no doubt that there are gas-bearing structures in the Berezovka district associated with an area of intensive development of the disjunctive dislocations in the Epihercynian basement that exist here, and also in the open zone of the Polar Urals with their northeastward trends. If one is guided by the map, one would expect to find the continuation of the Berezovka gas-bearing structural zone in the region to the northeast, beyond the Ob' River. On the other hand, the structures that run parallel to the Berezovka structures but are located further to the southeast may, perhaps, be more likely to be oil-bearing, since they occur in an area of deeper subsidence of the Paleozoic basement.

In the remaining territory of the Lowland, if one judges by the attached map, interesting prospects for oil and gas might be found in the northwestward-trending tectonic lines along the middle reaches of the Ob' and Irtysh Rivers. The zone of the Laryak-Korlykov step also deserves attention, since it contains a combination of the necessary structural and facies conditions.


100 0 100 200 300км
66
78
90
102
Dikson
Ust'-Port
Noril'sk
Baydaratsk depression
Салехард
Березово
Бурмантово
Кадель
Серов
Шаим
Леуши
Туринск
Sverdlovsk
Тюмень
Челябинск
Кустанай
Тобольск
Ханты-Мансийск
Сургут
Покур
Васюган
Тара
Викулово
Петропавловск
Омск
Купино
Барабинск
Новосибирск
Камень
Томск
Пудино
Ларьян
Корлики
Туруханск
Нижняя Тунгуска
Подкаменная Тунгуска
Kassk depression
Ангара
Ачинск
Krasnoyarsk
North Minusinsk depression
Восточный Саян
Абакан
Kuznetsk depression
Сталинск
Бийск
Горно-Алтайск
Kokchetav
Pavlodar
Akmolinsk
Semipalatinsk
Anuysk-Chuysk synclinorium
Terektinsk anticlinorium
Kalbinsk synclinorium
Charsk anticlinorium
Chingiz anticlinorium
Зайсан
68
60
52

Tectonic sketch map of the surface of the folded basement rocks of the Western Siberian shield:

- boundary between adjacent deposits of the platform mantle; 2 - structure contour lines of the irface of the folded basement (500 m section); 3 - same, inferred; 4 - same, supplementary (at 0 and 200 m intervals); 5 - domical and brachyanticlinal structures of the mantle and local proctions of the basement, as revealed by geophysical and drilling methods; 6 - extensive faults ineaments) primarily at the contacts between tectonic zones of different ages (tectonic sutures); - established; b - inferred; 7 - faults parallel to the main trend of the folding, associated th the development of the folded structures of this zone; a - established; b - inferred; 8 - ults cutting across the main trend of the folded zones; a - established; b - inferred; 9 - deessions filled with Paleozoic deposits (of the Minusinsk type); 10 - basins and grabens filled th extrusive-sedimentary formations of Permo-Triassic age; 11 - basins and grabens filled with aetian-Liassic and younger Mesozoic and Cenozoic deposits; 12 - area of supposed distribution the Lower and Middle Jurassic marine deposits; 13 - inferred boundary of the area of Upper irassic (Valanginian) marine deposits.

According to N. S. Shatskiy [4] and other vestigators, within the Russian platform ere is also a regular disposition of local ructures (placanticlines), in combination ith the main tectonic lines of the overall structural plan of the platform; in this sense one may scern some features of similarity in development between ancient pre-Paleozoic platforms id younger Epipaleozoic platforms.

Thus the general features of the structure the Western Siberian platform have been t forth here, with an eye to their possible actical use. It must be stressed, in conusion, that this article refers not to the ructure of separate small areas, but to the erall tectonic plan of the territory in question, epared according to definite views of the ructure and development of the major elements of the earth's crust, such as the Western berian platform as a whole; this plan may rve as a background for discovering the dissiton of concrete local structures.

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THE THRUST IN THE AREA OF THE KENEBEK-ZHONDYTAU MOUNTAINS, CENTRAL KAZAKHSTAN[5]

by

A. V. Luk'yanov, and I. G. Shcherba

Central Kazakhstan may be thought of as a block-folded region, broken by numerous faults along with the individual blocks have been displaced vertically relative to each other [2, 5]. By now, however, considerable information has been gathered that testifies to the great role

[5]Nadvig v rayone gor kenebek-zhondytau. Tsentral'nyy Kazakhstan, (pp. 105-109).

played by horizontal displacements along the faults in Kazakhstan [6]. These horizontal displacements have taken the forms of strike-slip movements, thrusts and movements opening fractures. The diagnostic identification of these structures is very complicated. The simplest horizontal displacements are those involved in thrusts; nevertheless in Central Kazakhstan up to the present only a limited number of structures of this type have been discovered. Thus each new and reliably established finding of thrusts is of great importance for a proper understanding of the geologic structure of Kazakhstan.

The structure described in this article is located in the Aksoran-Akzhal' fault zone, some 30 km east of the station of Basag on the Karaganda Railway line. The Kenebek-Zhondytau mountains and their surroundings are composed of Upper Devonian and Lower Carboniferous rocks. The volcanogenic-sedimentary and sedimentary strata of this age here, as in the other parts of the Aksoran-Akzhal' zone, occur conformably with each other, without any gaps in the deposits or traces of erosion. At the base of the section through the Kenebek-Zhondytau mountains lie deposits of the Frasnian stage. These are tuffs and lavas of albitophyric composition and their overlying tuffaceous sandstones and sandstones. Upward in the section they grade into the limestone strata of the Famennian and Tournaisian stages. There are gradual transitions between all these strata. The limestone series begins with calcareous sandstones and sandy limestones, which rapidly grade into blocky argillaceous and siliceous limestones. The latter, in turn, are gradually replaced in the upper part of the series by pure bedded limestones. The section is topped by a thick series of limy siltstones containing fauna of Tournaisian age. Younger formations, as dated by fauna, are not found in the area.

In the southern part of the Kenebek-Zhondytau mountains, however, there are volcanogenic formations whose geologic position has prompted a number of investigators of the region, such as V. V. Donskikh [4] and Chabdarov to consider them Upper Paleozoic. The discovery of the thrust described here is connected with the problem of their age.

All the above-listed rocks in the southern part of the Kenebek-Zhondytau mountains form an anticlinal fold with an axis running east-west and a flexure dipping steeply westward. In moving westward from the core of the anticline, one traverses the section described above, coming into younger and younger formations; finally (see illustration) one encounters a heterogeneous series of tuffs, tuffaceous sandstones and conglomerates which appear directly after the siltstones. In the northern part of this area, the series contains limestones in contact with the siltstones. Like the siltstones, the limestones and tuffaceous sandstones dip westward and southwestward. Their dip angles are usually on the order of 20 to 30°; the siltstones dip beneath the limestones and tuffs at an angle of 40 to 45°.

The boundary between these rocks and the siltstones in general repeats the configuration of the underlying limestone beds. Along the contact there is a small amount of fracturing, which is masked by the rock detritus that covers the bedrock outcrops.

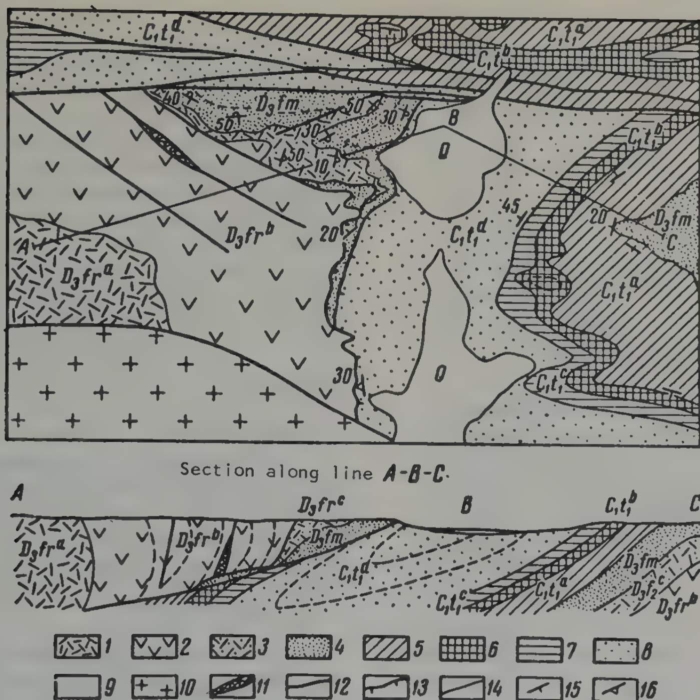
The clear occurrence of the tuffs and tuffaceous sandstones upon the siltstones, and the configuration of the outcrops of these strata, caused the above-mentioned geologists to consider the volcanogenic series to be a young formation (Viséan or Upper Paleozoic) unconformably overlying the Tournaisian siltstones. Within the siltstone series were included the limestones occurring at the contact with the tuffs.

In 1956-1957, however, in the limestones "capping" the siltstones, N. A. Pupyshev and the present writers collected a Lower Famennian fauna. This overthrew the basis for thinking that the extrusives were of fairly recent age, all the more because in the adjacent areas no port-Tournaisian extrusives are known, and the Visean stage is represented by sedimentary marine formations [1, 3]. N. A. Pupyshev, by analogy with the surrounding regions, assigned the tuffs to the Frasnian stage, considering that they are separated from the siltstones by a fault and that the Famennian limestones crop out in tectonic lenses. This interpretation of their age fitted in well with the conceptions of the geology of the region. Only the character of the displacement along the fault and the nature of the very gently sloping contact between the limestones and siltstones remained unexplained.

In the summer of 1959 the present writers studied this structure in detail, with its flat-lying thrust and intensively overturned beds in the allochthone. The internal structure of the tuff field (the allochthone), the position of the limestones within it, and the nature of the contacts between all the types of rocks were determined.

As already noted earlier, at the contact with the siltstone series in the west there are various rocks: limestones and various tuffs. But this contact is everywhere very sharp, and dips gently westward.

In the northern part of the area the siltstones form a contact with the limestones. The contact is well marked on the eastern slope of a small volcanic cone: the base of the cone is composed of siltstones which are highly



Geologic sketch map and section of the southern part of the Kenebek-Zhondytau Mountains

Frasnian stage: 1 - quartz porphyry tuffs; 2 - albitophyre tuffs; 3 - tuffaceous sandstones. Famennian stage: 4 - sandy limestones. Tournaisian stage: 5 - blocky limestones; 6 - nodular bedded red limestones; 7 - clastic-crystalline limestones; 8 - siltstones; 9 - Quaternary deposits; 10 - granites; 11 - tectonite. Disjunctive and faulted dislocations: 12 - normal and normal strike-slip faults; 13 - thrusts; 14 - stratigraphic contacts; 15 - strike and dip symbols of beds; 16 - strike and dip symbols for slickensided and fault surfaces.

ractured near the contact and broken up into ngular blocky fragments. Their mode of occurrence was impossible to determine. Limetones are exposed at the tip of the cone. The imestone-siltstone contact is inclined westvard at an angle of about 20 to 30°. In the icinity of the contact the limestones are roken by numerous small fractures and are ecrystallized. The limestone beds are almost arallel to the contact with the siltstones and ip westward at 30°. Farther south one oberves the same picture, and the bedding of the imestones remains generally parallel to the ontact, but the width of their outcrop is harply decreased. Finally, there are gaps etween the limestone outcrops; in these gaps uffs enter into contact with the siltstones, and he limestones are disposed along this contact s thin, very limited lenses. The lenses radually decrease in size from north to south, nd the lenses themselves occur more sparsely n the same direction. There is also a change n the degree of alteration of the rocks, from resh limestones shattered only at the contact o marbles that are friable at the surfaces of ll the lenses.

In the south the limestone-siltstone contact is intersected by numerous small gullies and rain channels. Along the bottoms of these, siltstone lenses are exposed beneath the tuffs, again stressing the flat-lying nature of the contact in the west. The limestone lenses cut through by a gully do not increase in thickness at its bottom, but like the contact they are inclined westward, beneath the tuffs.

The contact alteration in the tuffs and siltstones amounts mainly to fracturing, extending no more than 5 to 8 m from the contact. Within this interval the siltstones are usually broken by many small fractures filled with calcite. Sometimes they are convereted to breccia, also cemented with calcite. The same picture is to be seen near the contact in the tuffs as well; moreover in this zone the siltstones are silicified and the tuffs are leached. In the topographic depressions the leached tuffs extend as tongues along the gullies into the depths of the mass of rock.

All the facts cited here suggest the tectonic nature of the contact and its gently sloping westward dip. The magnitude of this slope can,

moreover, be judged from the slip surfaces in the siltstones about 100 m from their contact with a granitic intrusive (see illustration). These are smooth, polished surfaces of a plate that is regularly inclined westward at angles of 20 to 30°. Most frequently the angle of dip is 25°. This is evidently also the angle of inclination of the thrust sheet, or at least its upper part.

Thus the tuffs, tuffaceous sandstones and limestones occur in the allochthone of a thrust inclined toward the west. What are the relationships between them, and what is the internal structure of the allochthone?

Study of the contact between the limestones and the tuffaceous rocks in the southern part of the area leads one to conclude that the contact is tectonic; moreover it, too, is inclined westward and disappears gently beneath the tuffs. But farther north, the nature of the contact differs sharply. Here the limestones form two great outcrops, at whose edges one may observe a normal contact between the limestones and the tuffaceous sandstones with one rock gradually grading into the other. Here the section is as follows (from northeast to southwest):

1. Gray-brown sandy limestones with interbeds of crinoidal limestones dipping southwestward at 40°. There is fauna typical of the Kal'karatusov beds. The visible thickness is 50 m.

2. Medium-grained calcareous sandstones. 10 m.

3. Dense greenish-gray, fine-grained limy sandstones and thin-bedded green siltstones. 18 m.

4. The same limy sandstones with interbeds of bright-red tuffaceous sandstones and tuffites, whose number increases farther along in the section. 5 m.

5. Tuffaceous fine-grained sandstones and green tuffites. 25 m.

6. Tuffaceous conglomerate with poorly rounded and unsorted pebbles. 50 m.

7. Green and violet-colored agglomerates. 20 m.

8. These are followed by a series of greenish coarse-grained crystalline-clastic tuffs of albitophyres, of dense and massive structure. Among the tuffs are lavas, but the stratum in general is very homogeneous and unbedded.

Thus there is no doubt of the normal contact between the limestones and the tuffaceous sandstones and the gradual transition between these series. Wherever the contact is visible, the limestones dip beneath the tuffaceous sandstones (at angles of 30 to 50°), and the contact maintains the same attitude. Away from the contact the latter rocks sometimes dip more gently, at angles of 10 to 20°, and form a series of small folds. In a number of places it has been possible to observe tuffs lying upon the sandstones and tuffaceous sandstones (particularly in the above-described section). But in the majority of cases their contact follows the northwestward trending faults and the sandstone beds are cut at a sharp angle. It may be noted that in one of the faults cutting the tuffaceous strata there is a large, smoothed out lens of limestone (150 x 30 m) altered to tectonic breccia.

The tuff series is monotonous and unbedded. Two packets may be distinguished within it: green albitophyric tuffs and gray tuff-breccias of quartz porphyries, which occur in the southwestern part of the tuff field. These rocks replace each other very gradually. It has not, however, been possible to establish the sequence of their stratification.

Thus the following succession of beds has been established in the allochthone: limestones with fauna of the Kal'karatusov beds of the Famennian stage — sandstones and tuffaceous sandstones — tuffs with interlayered lavas. Nowhere else, in any of the adjacent areas of the Atasu-Zhamshinsk interfluve is such a section to be seen. The Famennian deposits always form a single thick series of limestones with the Tournaisian deposits; the base of this series contains fauna of the Kal'karatusov beds. At the same time, sandstones and tuffs underlie the limestones. Thus the triad of rocks as described in the allochthone is very typical of this area, but is always encountered in reverse sequence: tuffs, sandstones and limestones. In particular, a section which in its smallest details closely resembles the one described above is found about 1.5 km north of the thrust. There the tuffaceous strata underlie the Kal'karatusov beds and belong to the Frasnian stage.

The occurrence, on top of the Kal'karatusov beds, of a thick packet of tuffaceous deposits, connected to the former by a gradual transition is so far from agreeing with the sections through the Aksoran-Akzhal' zone that it cannot be accepted on the basis of observations made only over a small area. It appears that the rocks forming the allochthone occur in overturned succession.

To determine the tops and bottoms of the sections, the present authors have analyzed the composition of the pebbles in the conglomerates and the bedding in the sandstones. In the conglomerates lying within the allochthone, above the limestones and below the tuffs, there is a complete absence of limestone pebbles and a great abundance of pebbles of tuffs and

trusives which closely resemble those oc-rring "above" in the section. In the sand-ones, among the possible criteria for the ccession of bedding, there was found only a clical alternation of coarse-grained and fine-ained varieties. The cycle of alternation is me 4 to 5 cm thick. Each cycle begins with coarse-grained sandstone, which is gradually placed by medium-grained, fine-grained, etc. the end of the cycle there is a thin interlayer siltstone. This is followed by a sharp bound-y and the beginning of a new cycle. In every se, the coarse-grained part of the section s above the fine-grained.

On the basis of what has been presented, the esent writers believe that the allochthone has en overturned, and that the tuffaceous rocks long to the Frasnian stage. The amount of erturning of the beds is very great, as much 130 to 160° and in some cases even 170° — at is, the overturned beds cip at angles of) to 20° and even 10°. It appears that the beds ere overturned as a result of their downward nding along the surface of the thrust. The agnitude of the overturning decreases from st to west — in the direction away from the rust plane (illustration). Thus the thrust s a submeridional trend; it extends for 1.5 n and the amplitude of its displacement is less than 800 to 900 m.

On the north and south this structure is unded by faults of the normal or normal-rike-slip type, with a considerably greater tent and amplitude of displacement. These ults are the chief structure of the Aksoran-kzhal' fault zone. The thrust fault, which so forms part of this zone as an element of s structure, is considerably less typical of (a second thrust, with a steeper thrust plane, as found by the present authors between enebek-Zhondytau and Zhil'dytau Mountains). he discovery of these thrust faults greatly ex-ands the conception of the morphology of fault ones and underlines the complexity of the echanics of their formation, which is evidently ue to various stresses acting in different di-ections.

It may be noted that gently sloping thrusts ave been found in other areas of the Atasu-Mointa interfluve, by Ye. I. Sizova and V. D. oznesenskiy near the village of Novyy Khutor nd by Ye. B. Al'perovich and A. P. Semenova the area of Mt. Zhilan. M. P. Rusakov [7] nd A. I. Suvorov [8] have written about the rusts of Central Kazakhstan. There is every eason to support that, with detailed mapping f the territory of Central Kazakhstan, thrusts ill also be found in many other places of this egion.

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REVIEWS AND DISCUSSIONS

SOME CRITICAL REMARKS ON THE ARTICLE BY KONRAD BENES ENTITLED "PALEOMYCOLOGY — A NEW TREND IN THE MICROSCOPIC INVESTIGATION OF COALS"[1, 2]

by

A. A. Larishchev

This article is devoted to the interesting, but far from new branch of paleobotany — the study of the remains of fungi in fossil coals. At the beginning of the article we read: "In the field of paleobotanical studies up to the present time no attention has been paid to the fact that the substance of coals contains the carbonized remains of fossilized moulds and fungi". Even if we limit ourselves to paleomycological papers in Russian and other languages that are devoted to the remains of fungi in coals alone, and exclude the others, such as the paper by I. Felix, 1894, Beneš' conclusion will still appear no more convincing.

Below we shall attempt, from particular examples, to show that fossil fungi remains from the coals of the Soviet Union have long been known and published in photographs; they have been described and studied both as a method of coal petrography — that is, in thin sections and polished sections — and as maceration products — as a method of palynology. Moreover the unqualified advantages of the palynological method, by which one can examine these remains in isolation, in a viscous medium, by direct observation of the object itself from various sides, have been revealed beyond any measure of doubt.

Still more valuable results can be obtained by the combined use of both method together. Beneš' assertion that the basis of palynology must be the study of the remains in polished sections of coal is not true. In polished and thin sections the investigator is confronted with random, and extremely varied, cross sections sometimes through the same fossil remains; the study of these sections alone is far from sufficient, as will be quite obvious from the paleobotanical standpoint. Another oversight on the part of the author of this paper is the absence in his list of references of any literature published in the Russian language on this problem.

As early as 1915 the well known Russian paleobotanist, M. D. Zalesskiy, in his remarkable paper on "The Natural History of a Single Coal", which was devoted to a petrographic study of sapromixite, one of several varieties of Devonian coals of the Barzass type, carefully described and presented microphotographs of a number of representatives of fossil moulds and fungi. In addition to his detailed identifications, the author attempted, by means of botanical comparison and analyses of the Latin names, to produce a terminology for fossil remains that would be as close as possible to the artifical and formal nomenclature of botany, which the majority of coal petrographers had adopted in studying these remains only from sections. These coal petrographers included many non-Russign scientists: Z. Stach, V. Picard, R. Potonié, K. Beneš and others. Single thin sections, in the overwhelming majority of investigations, cannot lead to the best and, if one may use the expression, most "botanical" results. It should also be noted that M. D. Zalesskiy's work was accompanied by a complete translation of the entire text into French, instead of a mere abstract, as is most often the case with other authors.

Excellent, clear photographs of remains of fungi in both high and low relief in thin sections of Permian and Carboniferous coals, attempts at classification, descriptions and various explanations of their nature and genesis are to be found in published studies by I. I. Ammosov [1], in manuals and tests by Yu. A. Zhemchuzhnikov (1934 and 1935), in a monograph by Zhemchuzhnikov on the general geology of fossil coals (1948), in papers by A. A. Larishchev [2-5] and

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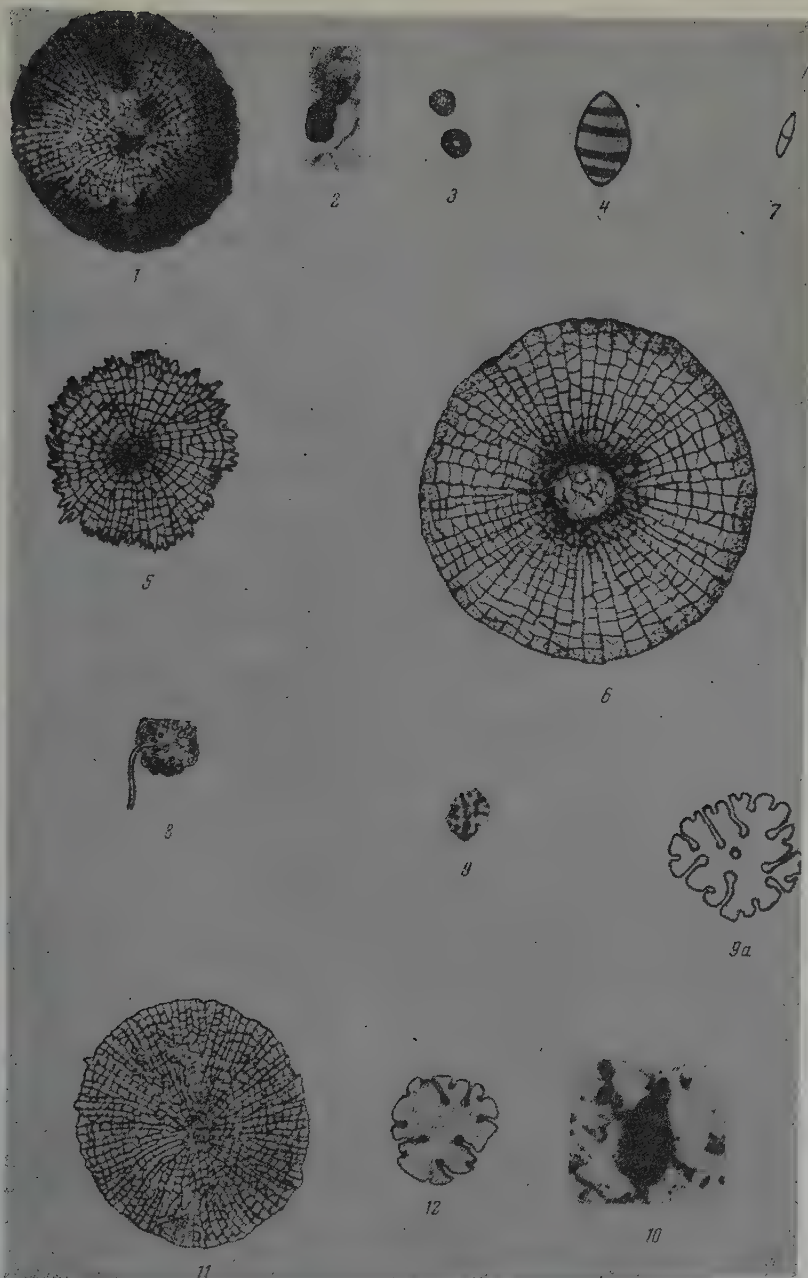
[1]Nekotoryye kriticheskiye zamechaniya k stat'ye konrada benesha "paleomikologiya—novoye napravleniye mikroskopicheskikh issledovaniy ugley", (pp. 110-112).

[2]Translated from the Czech paper of Yu. V. Shuf, Izvestiya Akad. Nauk S. S. S. R., Seriya Geologicheskaya, No. 11, 1960. The editors have also received a letter from N. Drozdova (Carbonometric Laboratory of the Scientific-Research Institute for the Geology of the Arctic), which points out some inaccuracies in the translation of this article.



nany other sources; they may also be seen nanuscripts and field reports.

There can be no doubt also of the value of dies carried out in the U. S. S. R. of fossil gi remains in coals as products of maceration hat is, studies using the spore-pollen meth- In 1953 was published the first work in this d by A. A. Chiguryayeva [6], which describ- and illustrates very numerous remains of rtiary fungi from various clays and other imentary rocks in the European part of the Soviet Union. The superb microphotographs of the remains of fungi, show that almost all these forms, with very few exceptions, have been observed by the present writer in maceration products of Tertiary coals from numerous locations in Western Siberia. In 1953 the present writer presented a report at the scientific conference at Tomsk University and in 1956 published a paper on certain rarer remains of fungi, which nevertheless are of more precise biostratigraphic significance, from the Eocene deposits of the Northern Hemisphere [5]. The question of the

narrow chronological position of Phragmothyrites eocaenica Edwards may perhaps sometime be more closely defined or expanded, but up to this time there are no known data on any other stratigraphic occurrence of this specific and characteristic fungus fossil [5].

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CHRONICLE

REE LECTURES BY PROFESSOR H. H. READ ON "A GENERAL SYNTHESIS OF THE LEDONIAN METAMORPHISM, PLUTONISM AND OROGENY IN GREAT BRITAIN"[1]

by

A. P. Lebedev, and Ye. V. Pavlovskiy

n March of 1961 the Soviet Union was host to of the most distinguished present-day lish scientists, the geologist and petro-pher H. H. Read, Fellow of the Royal Society, e-president of the Geological Society of don and Professor at the Imperial College of nce and Technology of the University of don.

Professor Read is widely known both in Great ain and far beyond for his numerous studies ncient metamorphic series and of the rocks he granite family. H. H. Read has made st of his own field investigations in the zone Caledonian structures in Ireland and Scotland. has devoted more than fifty years to the ly of these extremely interesting areas and continuing these studies at the present time.

H. H. Read's series of papers on the "granite blem", two of which have been published in sian translation in the U. S. S. R.,[2] are nous throughout the world. These papers, as well known, examine the various possible ans of formation of the granites in the earth's st, the "space problem" and the relationship ween the process of granitization and regional contact metamorphism, the phenomenon of gatization and folding. These papers will be particular interest to our readers. The ws developed by Professor Read of the existence of various genetic types of granites are also to a great degree applicable to the areas of development of the corresponding rocks in the U. S. S. R., composed of metamorphic rocks of the Pre-cambrian and Paleozoic.

The first lecture by Professor Read was devoted to an "Introduction to the Precambrian and Caledonian Geology of the British Isles".[3]

Northwestern Scotland and the adjacent islands contain ancient Precambrian formations — the Lewisian complex of gneisses, which have undergone metamorphism twice (the older stage of metamorphism was the Scowrie, 2460 to 2700 million years ago; the later was the Laxfordian, 1600 to 1400 million years ago). The Lewisian complex contains various intrusives of acidic and basic composition. The deeply eroded folded basement of the Lewisian is transgressively and unconformably overlain by the thick, non-metamorphosed terrigenous strata of the Torridonian, which in turn is unconformably covered by a transgressive series of the Cambrian and Lower Ordovician. Over all these formations of different ages within the Moine thrust zone the folded complex of the Moine series moved from the southeast; the Moine folded series is particularly extensive in the Northern and Grampian uplands of Scotland.

The Moine series consists of intensively metamorphosed terrigenous rocks — sandstones, marls and clay shales. The folded structures in

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[1]Tri lektsii professora G. G. Rida "Obshchiy sintez ledonskogo metamorfizma, plutonizma i organii v likobritanii", (pp. 123-128).

[2]"Reflections on Granite. Granite and Granites." the collection, "Problems of Granite Formation": latel'stvo Inostrannoy Literatury (Foreign Litera-e Press, 1949-1950.

[3]See also: H. H. Read, "The last twenty years work in the Moine series of Scotland", Verh. Ned. Geol. Mi jnb. Genoot., 16, 1956; J. Phemister, "British regional geology — Scotland: the Northern Highlands", Geol. Surv. and Museum, Edinburgh, 1960; J. Sutton, "Some structural problems in the Scottish Highlands", Rep. of the lst sess. I. G. C., p. 18, Copenhagen, 1960; H. et G. Termier, "L'Evolution de la lithosphere", Paris, 1956-1957; Ye. V. Pavlovskiy, "Some features of the Caledonian structures of Scotland", Izv. Akad. Nauk S. S. S. R., Ser. Geol., No. 6-7, 1957; W. Schreyer, "Geologisch-Petrographische Beobachtungen in der Schottischen Highlands", Geol. Rundsch., Vol. 46, 1957.

the Moine series are extremely complicated. Two systems of folds of different ages may be distinguished here. In the Grampian upland the Moine series is conformably overlain by the younger Dalradian strata. These include limestones, quartzites, a layer of glaciomarine deposits (the "boulder stratum"), clay rocks and graywackes which in places show excellent cross, sorted and cyclical bedding. The upper part of the section through the Dalradian series has abundant apilites and their tuffs. The Dalradian series is characterized by complex folded structure, with an abundance of recumbent folds of various magnitudes, which sometimes form entire "cascades". The rocks of the Moine series were probably laid down in the Precambrian; they have undergone metamorphism two times (700 and 400 million years ago respectively).

The Grampian upland, which is made up of the Moine and Dalradian series, has extensive manifestations of regional metamorphism. It is here that the metamorphic zones of the Barrovian were distinguished: the chlorite, biotite, garnet, staurolite, disthene and, finally, sillimanite zones. The igneous activity was in some cases later than the folding. Zones of the Buchan type contain andalusite-cordierite schists associated with the rise of the migmatization front on the periphery of domical uplifts.

After the completion of the Early Caledonian (pre-Ordovician) folding new geosynclines developed, in which there was an accumulation of thick (up to 10 km) series of graywackes and rocks of the ophiolitic series; these ultimately went through the Late Caledonian orogeny (end of the Silurian). Thereafter many intrusives were injected, particularly the large gabbroidal masses in Scotland, which are probably associated with deep faults, and still later were formed large intrusive bodies of granites which were violently irrupted into the surrounding rocks. The granite intrusives are accompanied by numerous lamprophyre bodies.

Devonian sedimentary and volcanogenic (andesites) formations lie unconformably upon the eroded surface of the folded Caledonian complex. In the Devonian new granite bodies appeared which, in H. H. Read's opinion, belong to the category of "permissible" intrusions which were injected "permissibly" into the host rocks during the formation of ring-shaped normal faults, caldera-like depressions and other structures creating favorable conditions for the passive intrusion of the granites. These "permissible" intrusives are, so to speak, the products of volcanism. They differ sharply from the granites of the first group (which were forcibly intruded) and are not accompanied by lamprophyres.

H. H. Read's second lecture was on the problem of the formation of the Dalradian granites in Ireland. The main object of his investigat[...] was the area of Donegal, in which granites o[f] various types are developed.

In the Donegal area, which is cut by a larg[e] fault (a continuation of the Great Glen fault), [...] genetic types of granites are extremely varie[d] because of the different conditions of their for[-]mation, as well as many different compositio[n] of the metamorphosed host rocks and their str[uc]tures. Up to the moment of intrusion of the granites, the surrounding rocks had been very slightly metamorphosed, to the degree of the green slate facies. All their succeeding meta[-]morphism was due to the action of the granites.

The granites were intruded as the result of various processes, the chief among which wer[e] 1) "reaction" injection, 2) active ("forcible") injection and 3) passive ("permissible") injection.

The first type (the Thorr granites) includes the cases in which the granite material reacte[d] with the surrounding rocks and in places meta[-]somatically replaced them. In their replacement of the host rocks the granites preserved relicts of the latters' initial tectonic structure which Professor Read calls phantom or "illus[ory"] stratigraphy. There was intensive reaction be[-]tween the granite "juice" ("ichor" or magma) and the surrounding rocks. Around the massi[f] formed in this manner appeared aureoles of andalusite hornstones, which were later almost completely replaced by muscovite-fibrol[ite] schists. Here the mechanism of "granitizatio[n"] is not very clear.

The actively intruded granites are those of the Ardar massif. They are surrounded by a series of small bodies of appinites (a variety of lamprophyres), consisting of olivine, pyroxene, hornblende and alkali feldspar and shot through with cylindrical, pipe-shaped cavities (diatremes). The appinites are always associ[-]ated with the granites of the second group, but they are older than the granites. The Ardar pluton is composed of tonalites on the periphe[ry,] of partially sheared diorites in the middle and of a granite core at the very center. Apparen[tly] the marginal parts are the earlier, after whic[h] newer and newer portions of granite magma entered and expanded the massif from within. Inasmuch as the diorites and tonalites are tor[n] apart into boudinage structures, within the su[r]rounding rocks (represented by andalusite schists) there took place a complex periphera[l] folding complicated by strike-slips and thrust[s.] The degree of metasomatism here is less tha[n] around the granites of the first type.

The granites of this group also include thos[e] of the "main massif" of Donegal, which is 48 [km] long and 6 to 8 km wide. A great part of this massif is covered by the roof rocks. The ma[s]sif shows clear banding and traces of flow

uctures. The banding (linear elongated nes of dark inclusions) may be due to schlieren reworked xenoliths. The xenoliths were aprently moved to distances of some 30 km. e magma here moved southwestward, almost rizontally, as shown by a study of the strucres. Linear structures are also to be seen the host rocks; these also arose through the w of the granite magma. The surrounding ystalline schists contain kyanite, staurolite, lorite, biotite, muscovite and some andalusite. ere a study of thin sections is helpful in drmining the relationships between crystalzation and deformation. Typically, in conast to the host rocks, the xenoliths show no aty foliation.

An example of the granites formed by passive jection or, as H. H. Read picturesquely puts , intruded "with the permission of the host cks" are the granites of the Rosses massif. ese form a ring-shaped complex consisting four "phases" successively intruded along a ng structure created by caldera-like subdence. These granites are not accompanied any thermal metamorphism. The Rosses ng complex was intruded into the Thorr anodiorites.

Characteristically, each of the four granite hases" was intruded into the contact aureole the preceding phase, so that later metaorphism was also superimposed upon earlier. he inner ring was composed of the earliest ranite "phase".

Some regularities have been observed in the istribution of the aluminosilicates in the conct aureoles. Apparently the sillimanite in the onegal is always metasomatic, the andalusite always associated with the outer and the yanite with the inner contact zone. Thus the ndalusite is probably a lower-temperature ineral than the hyanite. But additional studies re needed for a complete solution to these uestions.

The third lecture set forth a "general synhesis of the Caledonian plutonism and orogeny n Great Britain". H. H. Read began his talk vith the remark that "the nature of the Caledonan structures is perfectly clear to everyone xcept those who have worked in Scotland". In ead's view, the Caledonian structures of cotland may be divided into two zones: 1) metanorphic and 2) non-metamorphic. In the metanorphic zone the sedimentary rocks are represented by the Moine and Dalradian series, which re probably of Late Precambrian origin. Their netamorphism, migmatization and folding are f pre-Arenigian age. These two series form complex which Read calls the Early Caledonan.

The zone of non-metamorphosed rocks is epresented partly in Scotland, Ireland and England and most of all in Wales. The Cambrian, Ordovician and Silurian deposits in Wales were pressed into folds in post-Silurian times; they are only slightly metamorphosed. The last episode of Caledonian history – the Old Red Sandstone of the Devonian – is represented by deposits of molasse type.

The Early Caledonian (pre-Arenigian) orogeny produced the metamorphism and folding of the Precambrian sediments. With these are also associated the alkaline intrusives that penetrated them along fault zones. Their precise age is unknown. They were followed by the successive intrusion of the gabbroic massifs in Northeastern Scotland.

The non-metamorphic Caledonian zone was involved in the Late Caledonian folding (at the end of the Silurian), after which dolerite was intruded. The last events in both Caledonian zones ran parallel courses: in both there appeared an abundance of appinite bodies, followed by the "forcible" intrusion of granitoids. All this took place in pre-Devonian times. With the Devonian volcanism and molasse sedimentation are associated the "latest" granites, which formed ring-shaped intrusives.

The manifestations of igneous activity in Moinian and Dalradian times are worthy of attention. The Moine and Dalradian series were crumpled into folds and metamorphosed before the Arenigian age (the Early Caledonian orogeny). The Moine series sediments were deposited under deltaic conditions, those of the Lower Dalradian under shelf conditions and those of the Upper Dalradian in a eugeosyncline. In Northern Scotland the Early Caledonian orogeny was accompanied by ophiolitic intrusions and spilitic lava outpourings. H. H. Read believes that the same northern part of Scotland also contains ancient Precambrian granite intrusives that were injected into the rocks of the Moine series and produced aureoles of contact metamorphism within the latter. Only after all these events did regional metamorphism take place. Elements of ancient pre-Caledonian igneous activity also occur on the island of Anglesey, where there is a complete granite series consisting of migmatites, "permissible" granites and appinites.

In the zone of non-metamorphosed Caledonian rocks, as in Wales, where a typical geosynclinal regime reigned during the Early Paleozoic, there are no traces whatever of ophiolitic migmatism. The Ordovician volcanogenic series are composed principally of rhyolites. In the Caledonian metamorphic zone, after various manifestations of older magmatic activity, the Early Caledonian migmatization, metamorphism and folding took place. In considering migmatization, one must distinguish between regional and local migmatites. Local migmatites are developed only around certain

actively injected ("forcible") granite intrusives. They do not represent any considerable event in the development of the given part of the earth's crust, as contrasted with regional migmatites, which are sometimes spread over areas (for instance around Saterland, Northern Scotland) of hundreds of thousands of square kilometers.

The first event in the Early Caledonian history of the metamorphic zone was the regional migmatization. The next important stage was the intrusion of large amounts of gabbroic magma. Of particular interest is the great gabbro Inch massif in Scotland. Here the magma at many points reacted with the host rocks of pelitic composition, to form cordierite-containing contamination products with rhombic pyroxenes (cordierite norites). The Inch massif includes an entire differentiated series of rocks — dunites, olivinites, fayalite and orthoclase-olivine gabbros, and leucocratic syenites.

The rise of the basic magma was accompanied by a rise of the migmatization front to a very high level. Simultaneously folding took place. All these events were apparently closely connected.

Thereafter numerous small bodies of appinites were intruded. Sometimes these were intrusive breccias consisting of rock fragments occurring at depths of 4 km or more, but usually they were irrupted rocks of a composition similar to that of the lamprophyres that were mentioned earlier. Among these there are also olivine monzonites and sometimes rocks of ultrabasic composition. The appinites are closely associated with the actively intruded granites located on the peripheries of the granite bodies, but the appinites are always older than the granites. This association of the appinites with the "forcible" granite intrusives is very typical. The actively intruded granites are about 400 million years old.

The later (Devonian) post-orogenic granite ring intrusives are especially common in the southeastern part of Scotland. These appeared against the background of a general uplift of the earth's crust.

Study of the igneous formation leads one to the conclusion that their type may be a criterion for assigning a given part of the earth's crust to one structural stage or another. At the same time the phenomena of magmatism may be used to distinguish and group the events in the history of the earth's crust, although it must be stressed that the relationships between the processes of metamorphism, migmatization, igneous activity and orogeny in the Caledonian parts of Great Britain are exceedingly complicated and do not always fit into the standard schemes.

H. H. Read's lectures aroused great interest on the part of Soviet geologists and petrographers, especially among those who have been occupied in studying the areas of extensive occurrence of ancient metamorphic series. H. H. Read has attempted to produce an original explanation of the complicated chain of events that took place within the British Isles in the Late Precambrian and Early Paleozoic — an area that has long been considered the "classic" Caledonian region. Note should also be taken of the clear evolution of Read's views on the "granite problem", leading to a more purely magmatic interpretation of their formation than he held in his earlier days.

IN THE GEOLOGICAL INSTITUTE OF THE ACADEMY OF SCIENCES OF THE U. S. S. R. (ON THE OCCASION OF YE. V. PAVLOVSKIY'S SIXTIETH BIRTHDAY)[4]

by

A. M. Leytes, and M. S. Markov

On April 22, 1961 there was held a meeting of the Scientific Council of the Institute in honor of Professor Yevgeniy Vladimirovich Pavlovskiy, Doctor of Geological and Mineralogical Sciences, to mark his sixtieth birthday and the thirty-fifth year of his activities as a scientist and teacher.

The acting director of the Institute, P. P. Timofeyev, and Academician A. L. Yanskin tendered their congratulations in the name of the Institute. A. L. Yanshin remarked that Ye. V. Pavlovskiy, a member of the Geological Society of France and of the International Association for the Study of the Depths of the Earth's Crust, is a distinguished specialist in the problems of theoretical tectonics and in the field of the regional tectonics of Siberia and extra-Alpine Europe. His pen has produced more than a hundred scientific papers, the total amount of which exceeds that of more than 200 authors. A. L. Yanshin presented a brief review of Ye. V. Pavlovskiy's investigations and pointed out their great importance in the development of many major aspects and problems of the structure and history of the earth's crust (the process of arkogenesis, the particular features of the development of the crust in the Precambrian, the distinguishing of major structural units of the earth's crust such as zones of pericratonal subsidence, etc.). He especially stressed the daring, novelty and originality of the ideas in Pavlovskiy's papers, which are also replete with abundant and convincing factual material.

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[4]V Geologicheskom Institute Akad. Nauk S.S.S.R. (60-letiya Ye. V. Pavlovskogo), (p. 128).

Thereafter Ye. V. Pavlovskiy reported on the in conclusions of his work on "The Specific atures of the Early Stages of Development of Earth's Crust". The object of his investigation were the Canadian, Baltic, Ukrainian, Aldan and Anabar shield, as well as certain ancient structures of East and South Africa. His paper makes use of the results of current investigations of these shields carried out by geologists of many countries.

At the end of his report he was congratulated by the representatives of various institutions of the Academy of Sciences of the U. S. S. R. and various geological organizations. A long list was read of the many Soviet and foreign organizations and persons sending their greetings to Ye. V. Pavlovskiy.

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